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A Water Balance Evaluation of the Effects of Climate Variability and Human Modification on the Flow Regime of the Mississippi River, 1932-1988.

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A water balance evaluation of the effects of climate variability and human modification on the flow regime of the Mississippi River, 1932–1988

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A WATER BALANCE EVALUATION OF THE EFFECTS OF CLIMATE VARIABILITY AND HUMAN MODIFICATION ON THE FLOW REGIME OF THE MISSISSIPPI RIVER: 1932-1988

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ABSTRACT

Human activities affecting runoff and streamflow within the Mississippi River drainage basin have been and continue to be significant. Since the disastrous flood of 1927, the U.S. Army Corps of Engineers has established an extensive regulatory system for flood control purposes and navigational benefits. Along with these river system modifications, significant landscape alterations have occurred within the basin as a result of cultural and economical development. In light of the dramatic Midwest flood of 1993, it is increasingly necessary to understand the comprehensive relationship between streamflow, climatic variability, and human-induced change within the overall drainage basin.

A water balance methodology provides a framework for evaluating variations in measured discharge over time by separating the effects of land-use changes and river management from variations in streamflow caused by geographical and temporal variability in climate. Therefore, it is possible to evaluate the collective results of human modification within the watershed. This study utilizes Box-Jenkins statistical time series analysis to identify changes in the annual flow regime of the Mississippi River for water years 1932 to 1988. For comparative purposes, the Missouri and Ohio-Tennessee Rivers are also analyzed.

The water balance model explains as much as 96% of the variability in annual discharge within the Mississippi basin and the Missouri and Ohio-Tennessee subbasins. There is no evidence of statistically significant changes or long-term trends in annual precipitation, modelled runoff, and discharge in the Mississippi basin and the Missouri

and Ohio-Tennessee subbasins, but the annual discharge fractions generated by the Missouri and Ohio-Tennessee subbasins have changed significantly. This can be attributed to the fact that the size of the overall drainage basin is so great that even large-scale atmospheric circulation patterns cannot affect the entire drainage basin at the same time. The relationship between annual discharge and modelled runoff has not changed in the Mississippi and Missouri basins; however, the relationship between annual discharge and modelled runoff has changed significantly in the Ohio-Tennessee subbasin. This change can possibly be attributed to river management and land-use changes within the subbasin.

CHAPTER 1

INTRODUCTION AND LITERATURE REVIEW

1.0 The 1993 Flood

Flooding along the Mississippi and Missouri rivers during the summer of 1993 was stated to be the worst in the northern Mississippi River basin in over a century. Flood waters covered over 17 million acres in nine states, caused 48 deaths, and resulted in over $12 billion in losses (Lott 1993). In addition to normal climatic variability, the flooding of 1993 has been attributed to the following probable factors: an El Niño-Southern Oscillation (ENSO) event in 1992-1993 which resulted in persistent and excessive precipitation throughout the spring and summer of 1993; excessive winter snowpack in the Rocky Mountains; and an extremely wet fall in 1992 which resulted in saturated soil conditions in the Midwest (U.S. Department of Commerce 1993).

As the great flood of 1993 so dramatically demonstrated, major flood events within the Mississippi River drainage basin continue to occur, despite the concerted efforts of the U.S. Army Corps of Engineers to control and mitigate flooding. In addition to the probable causes for the 1993 flood listed above, the Corps of Engineers stands in the midst of a controversial debate about whether river management and regulation have, in fact, contributed to flood damages by overcontrolling the river. This debate extends further to the effects of land-use changes which have occurred in recent years. Wetland losses have removed areas which could have absorbed surplus water and thereby might have reduced the high river levels experienced in this flood (Midwestern Climate Center 1993).

Most often, development proceeds in a watershed without detailed information about how that development affects streamflow characteristics. Moreover, the resultant effects may not be apparent immediately. In this regard, Muller (1981) commented that:

> there is considerable evidence that we lack appreciation, understanding and wisdom of the overall interactions and consequences of land-use and engineering controls over entire river basins (p. 172).

Thus, the flood of 1993 raised the following questions which recur each time another dramatic climate event affects the Mississippi River system:

- Is excessive flooding or drought strictly a result of climate variability?;
- Have regulation and management of the river by the Corps of Engineers exacerbated the climatic effects?; and
- Have land-use changes played a role in compounding severe climatic effects?

This study investigates these questions by analyzing annual time series data for the Mississippi basin and evaluating changes that have occurred over the years 1932 to 1988.

1.1 Study Objectives

Considerable research has been carried out to evaluate the individual impacts of vegetation changes, river management, and urbanization on streamflow. However, it is less common and more difficult for researchers to examine the collective and cumulative effects of human actions on large river basins. Eagleson (1986) stated:

> Because of humanity's sheer numbers and its increasing capacity to affect large regions, the hydrologic cycle is being altered on a global scale with consequences for the human life support systems that are often counterintuitive. There is a growing need to assess comprehensively our agricultural, urban, and industrial activities, and to generate a body of knowledge on which to base plans for the future (p. 13S).

The actual response of streamflow to land-use changes depends on climate, geology, topography, soil, vegetation, and the area and type of land-use changes. A water balance methodology is used to separate the effects of human modification on streamflow from the effects of climate variability on streamflow, since discharge is affected by all of these factors, and the water balance runoff depends only on climatic input. Thus, the difference between measured discharge and modelled runoff (defined as the water balance residuals) provides information regarding the effects of physical changes within the drainage basin on streamflow.

The questions to be investigated in this study are:

1. How well does the water balance model predict annual streamflow for the Mississippi basin and its subbasins?
2. Has annual discharge from the overall basin and its subbasins changed over time? If so, has annual precipitation over the basin and its subbasins changed over time?
3. Have human modifications affected the annual streamflow regime in the Mississippi watershed and its subbasins?

In order to answer these questions, it is necessary to calculate water balance models for each drainage basin and compare the results with measured discharge data. Further, the time series data for precipitation, surplus, and discharge must be evaluated for significant changes over the period of study. Finally, the water balance residuals must be analyzed to determine if the effects of human activities can be detected. Therefore, the following steps comprise the major objectives of this study:

1. Develop working data sets for annual precipitation, surplus, modelled runoff, discharge, and water balance residuals for the overall Mississippi basin and selected subbasins;
2. Analyze discharge to evaluate hydrologic variability over the years studied;
3. Analyze precipitation, surplus, and modelled runoff to evaluate natural variability and the possibility of climatic change over the years studied; and
4. Analyze the water balance residuals for indications of human-induced change.

The results of this study should provide information about the effects of climatic variability and human activities on the annual Mississippi River flow regime. Potentially, this information can be used to optimize the future management of the Mississippi River system.

1.2 Historical Overview

Human activities affecting drainage and discharge within the Mississippi River drainage basin have been considerable. The construction of the first levee system began in 1717 with the founding of New Orleans, and as settlements developed along the river, the levee system was developed accordingly. By 1844, the levee system extended from below New Orleans to the Arkansas River on the west bank (except for a gap at Old River) and from below New Orleans to Baton Rouge on the east bank. Throughout this time period, flood control efforts existed almost entirely on a local level (U.S. Army Corps of Engineers 1974).

Federal involvement with the Mississippi River began in 1820, when federal funds were allocated for navigational surveys of the Mississippi and Ohio rivers. During this time, the Corps of Engineers was primarily concerned with navigational improvements (Koellner 1988). In 1849 and 1850, floods in the Lower Mississippi Valley directed national attention to the problem of flooding, yet an initial Congressional attempt at coordinated flood protection was unsuccessful. In 1879, the Mississippi River Commission (MRC) was established by Congress to coordinate navigation and flood control improvements. For many years thereafter, the MRC was authorized only to construct and repair levees and to improve and maintain the river channel for navigation. Flood control was not emphasized (U.S. Army Corps of Engineers 1974).

The first Flood Control Act (passed by Congress in 1917 as a result of the 1916 flood) and the second Flood Control Act (passed in 1923) served to clarify the jurisdiction of the MRC. However, it was only after the disastrous flood of 1927 that the federal government became formally committed to a prescribed program of flood control. The 1927 flood overtopped levees all along the river, caused 214 deaths, left nearly three-quarters of a million people homeless, and resulted in losses of $236 million dollars (Lower Mississippi Region Comprehensive Study Coordinating Committee 1974).

With the passage of the Flood Control Act of 1928, the Corps of Engineers began developing a vast program to control flooding in the Lower Mississippi River Valley by constructing reservoirs and levees. This legislation also directed the Corps of Engineers to evaluate flood control possibilities on Mississippi River tributaries. As a result, the federal flood control program expanded to include the Tennessee River on the basis that the river was a small yet critical factor in Mississippi River flood control (Roberts 1955).

In 1933, the Tennessee Valley Authority (TVA) was authorized (Droze 1965). The Flood Control Act of 1944 authorized the construction of Kentucky Dam on the Tennessee River (Roberts 1955) which resulted in the total control of discharge from the Tennessee River into the Ohio River.

Thus, federal policy rapidly accelerated river management and land-use changes within the drainage basin. During the ensuing years, the Corps of Engineers established an extensive regulatory system of levees, spillways, diversions, and reservoirs. In addition, channel improvement and stabilization measures, such as dikes, revetments, dredging, and foreshore protection, were installed for navigational benefits.

As a result of cultural and economical developments, significant landscape alterations also occurred within the drainage basin. The grassland and forested areas of the early 19th century were converted to intensively used agricultural lands. In the early 1930's, conservation programs were initiated throughout the Mississippi River basin to lessen soil loss rates. Trees and grasses were planted on denuded areas, small sediment-retention dams were constructed, streambanks were stabilized, and contour plowing methods were recommended. Over one-third of the basin is currently used for agricultural purposes (Keown, et al. 1986). In more recent years, additional lands have been converted to urbanized and industrialized areas.

1.3 Literature Review

This section presents a representative sample of the information available in the scientific literature concerning the analysis of streamflow variability, particularly in regard to climate change, the effects of human influences on streamflow, and related Mississippi River drainage basin research.

1.3.1 Streamflow Variability Studies

Bartlein (1982) used principal components analysis to examine the monthly flow records of 102 streams in the U.S. and southern Canada for the period 1950 to 1970. Stream selection was based on minimal alteration of natural flow. He found that most of the regional variability in discharge could be attributed to four basic large-scale anomaly patterns. The primary component corresponded to a pattern of below normal streamflow in most regions which reflected the widespread drought during the years 1953 to 1958. The remaining components corresponded to contrasting patterns of above and below normal streamflow in various regional areas. He concluded that these patterns were the result of climatic anomalies of relatively large-scale and long duration related to the general large-scale circulation of the atmosphere.

Lins (1985) also used a principal components analysis to evaluate spatial and temporal variability in streamflow in the U.S. over the years 1931 to 1978. Streamflow records were used for rivers with no reported regulation or diversion or where diversion amounted to less than 10% of the mean annual runoff. He identified five significant components of streamflow. Of these, the main pattern was representative of uniform conditions (either above or below normal discharge) across much of the U.S. with significant flooding or drought in the Middle Mississippi Valley. The study identified temporal patterns of streamflow across most of the U.S. as follows: normal to below normal flows occurred during the years 1931 to 1941; persistently above normal flows occurred from 1942 to 1952; significantly below normal flows resulted during the years 1953 to 1956; moderately below normal flows occurred from 1956 to 1968; and above normal flows predominated during the years 1969 to 1978.

Recent studies have attempted to evaluate the effects of the El Niño-Southern Oscillation (ENSO) phenomenon on streamflow. For example, Kahya and Dracup (1993) studied the relationship between ENSO years and unregulated streamflow over the period 1948 to 1988. ENSO years were identified as 1932, 1939, 1941, 1951, 1953, 1957, 1965, 1969, 1972, 1976, 1982, and 1986 (as based on Rasmusson and Carpenter 1983). They investigated extreme streamflow events (flooding and drought) with respect to ENSO years and identified four regions with significant ENSO-related streamflow signals. For the years studied, ENSO events produced wet conditions within the Gulf of Mexico and North Central regions and produced dry conditions within the Northeast and Pacific Northwest regions.

1.3.2 Effects of Climate Change on Streamflow

In recent years, climate change has been the focus of a great deal of research with respect to the effects of global climate change and global warming on streamflow. For example, Revelle and Waggoner (1983) utilized divisional climatic data and multiple regression techniques to predict the effects of climate change on the mean annual flow of the Colorado River. Discharge data for the Upper Colorado basin for the period 1931 to 1976 were adjusted to account for depletions within the basin, evaporation from reservoirs, and changes in reservoir storage. They estimated that a 2 °C increase in temperature would reduce flow by 29% $\pm$ 6%, and a 10% decrease in precipitation would further reduce flow by 11% $\pm$ 1.4%. Therefore, they recommended that hydrologists carefully consider possible climatic change when planning and constructing major water resource systems.

Meko and Stockton (1984) examined streamflow (with an emphasis on drought) in the western United States for the years 1914 to 1980 to define the natural variability of runoff and to determine whether climate-related changes occurred during those years. Twenty-six basins were studied which ranged in size from 1241 km^2 to 55,814 km^2. Mean annual streamflow in two regions showed evidence of nonstationarity when comparing the periods 1914-1946 to 1947-1980. A difference of means test showed that flow rates increased significantly in the Pacific Northwest (by 20%) and decreased significantly in the Upper Colorado (by 15% to 20%). They concluded that climatic variations caused significant hydrological changes in the western U.S. during this period.

Kite (1989) examined the mean monthly time series of lake levels (Great Salt Lake, Lake Victoria, and Lake Superior) and river flows for evidence of climatic change in the form of linear trends, periodicities, autoregressions, and random residuals. Flows were analyzed for three rivers in Canada over the years 1916 to 1986: the Bow River near Banff; the Saugeen River in Ontario; and St. Mary's River in Nova Scotia. His study failed to detect any evidence of climate change caused by the greenhouse effect.

Gleick (1986) discussed six limiting factors to be considered when selecting and implementing a hydrologic model for evaluating the effects of climate change as follows:

1. Model accuracy;
2. The degree to which model accuracy is based on existing climate conditions;
3. Availability of input data as well as sufficient historical data for comparative purposes;
4. Input data accuracy;
5. Model flexibility, ease of application, and adaptability to diverse climate and hydrologic conditions; and
6. Compatibility with existing general circulation models.

Frequently, a water balance methodology is used because it adequately meets these criteria. For example, Gleick (1987) used a water balance methodology to estimate the effects of climate change on the Sacramento basin for the years 1931 to 1980. This watershed has been affected by a number of large reservoirs and heavy irrigation withdrawals. A plot of monthly predicted and observed runoff demonstrated that the observed spring runoff occurred later in the year than predicted, and the observed summer runoff remained higher than predicted. This was determined to be a result of having no allowance for snowfall and snowmelt in the model. Runoff estimations were improved by dividing the basin into a two-basin model based on elevation differences. Gleick tested the accuracy of his model on annual, seasonal, and monthly bases. He also plotted residuals for each of the 50-year monthly averages to detect trends in model runoff which would have demonstrated a consistent bias in the model, and no trends were found.

Rowe, et al. (1991) used the Thornthwaite-Mather water balance to model streamflow in the Little Blue River above Fairbury, Nebraska, since 1900. They applied hypothetical climate changes to temperature and precipitation data and reran the water balance model. The resulting water balance estimations of streamflow were used to evaluate the sensitivity of streamflow to various changes in climate.

McCabe and Wolock (1992) studied the effects of climatic variability and change within the Delaware River basin by utilizing the Thornthwaite moisture index (Thornthwaite and Mather 1955). This index depends only on temperature and precipitation and is not affected by estimates of soil moisture capacity, soil moisture availability, or actual evapotranspiration. Observed mean annual moisture indices for the years 1950 to 1983 were compared to mean values calculated by global climate models for doubled CO_2 concentrations. They concluded that natural variability may mask long-term trends in hydrologic variables, thus delaying the detection of climate change by many years.

Matalas (1990) discussed the analysis of historical hydrologic records in order to detect evidence of climate change in recent decades. He emphasized that:

> Trends in the parameters may be induced by changes in factors other than climate. For example, land-use practices change with time, and those changes can and do affect the hydrology of a region. Thus, if there are trends in the parameters, it is difficult to say if the trends have been induced by change in climate or by change in land-use practice or, for that matter, by change in some other factor. There is the possibility that climate and land-use practice could both change more or less contemporaneously in a manner that would not induce trends in the parameters. Trends in the parameters are not necessarily evidence of climate change. No trend in the parameters is noninformative regarding climate change (Matalas 1990, p. 140).

1.3.3 Effects of Land-use Changes and River Management on Streamflow

An extensive body of research is available with respect to the effects of specific human actions on streamflow. Mrowka (1974) identified types of human impacts and summarized their corresponding effects on stream regimen as follows:

1. Direct channel manipulation (e.g., channelization, bank manipulation, and levee construction) results in a faster time of transmission through the channelized reach;
2. Dam and reservoir construction results in increased evaporative losses and significantly affects the downstream flow regime;

3. Irrigation diversions result in reduced downstream discharge;
4. Modification of watershed characteristics (e.g., alteration of vegetation, plowing, tilling, terracing, other cultivation practices, and various drainage practices) results in varied effects upon streamflow that may be competing; and
5. Urbanization (e.g., the construction of storm drains, concrete lining of channels, and covering of the watershed surface) results in reduced infiltration, reduced groundwater recharge, and therefore, reduced bed, bank, and floodplain storage.

Since the 1970's, research has been conducted in the Soviet Union to determine human impacts on annual runoff from vital river basins (Shiklomanov, et al. 1989). A model of the Volga River was developed in order to evaluate changes due to irrigation, municipal and industrial consumption, and evaporation from reservoir storage. Despite the fact that runoff processes are generally considered to be stationary, random processes, "it is hardly possible to find a water body the regime of which is not changed by man's activity in the most economically developed regions of the USSR (Shiklomanov, et al. 1989, p. 198)."

Hall, et al. (1989) created a model to simulate lake levels for Flathead Lake in Montana in order to aid 15 agencies in developing and managing realistic lake water levels and downstream flow rates. Kerr Dam (at the outlet of Flathead Lake) was constructed in 1937, and Hungry Horse Dam (on the South Fork of the Upper Flathead River) was completed in 1952. The authors determined that the flow of the Lower

Flathead River had been affected by regulation as follows: mean flows during spring runoff were 45% lower than pre-dam mean flows; and mean flows during winter increased 89% over pre-dam mean flows. Based on daily values of U.S. Geological Survey (USGS) streamflow data, precipitation data, lake level readings, calculated evaporation, and irrigation withdrawals for the period 1950 to 1986, their model was able to predict lake levels within a few centimeters. It was assumed that disagreement between the model and empirical lake levels was caused by inaccurate USGS flow data during periods of high discharge.

Water balance methods have been used frequently to analyze the effects of land-use changes on streamflow. Muller (1966) conducted the first study to use a water balance methodology to evaluate the effects of vegetation treatment on streamflow. He utilized the Thornthwaite water balance model as a hydroclimatological control to determine the effects of forest cover on water yield within four experimental watersheds in central New York. His study showed substantial reductions in water yield as a result of farm abandonment and partial reforestation.

Mather (1981) compared runoff calculated from the Thornthwaite water balance method with measured runoff for 28 small drainage basins in the northeastern U.S. Basin size varied from 7.3 km^2 to 813 km^2. The standard Thornthwaite water balance was modified to include an evaluation of effective precipitation based on overland runoff from intense rainfall using Soil Conservation Service (SCS) tables. Modelled runoff was calculated by using a lag factor of 25% for available surplus in a given month with the remaining 75% added to the available surplus for the following month.

These adjustments resulted in good agreement between calculated and measured runoff for both an annual and monthly basis. One drainage basin included in this study showed the effects of human interactions due to ditching operations and gauge modifications.

Shelton (1981) predicted the effects of accelerating forest and agricultural resource development on the quantity and timing of runoff in the Deschutes River basin (27,300 km^2) in Oregon. He utilized a general basin hydrologic cascade model based on a climatic water balance and two possible groundwater storage systems. Water balances were calculated on a monthly basis for the period 1952 to 1957. Effects of timber removal via clear-cutting were estimated by reducing soil moisture storage capacity to two inches over forested areas of the basin, which resulted in an average increase in runoff of 12%. The effects of irrigated agricultural expansion were estimated by assuming an increase in water diversions based on a maximum irrigated acreage, which resulted in an average decrease in runoff of 8.4%.

Shelton (1985) compared measured and simulated runoff for the Deschutes River basin for the years 1943 to 1956 using a spatially disaggregate moisture balance watershed model. Subsequent years were not included in the study because the natural flow of the river was disrupted by increased diversions for irrigation and construction of a major dam. He assumed that runoff models for watersheds larger than 1,000 km^2 produced inaccurate estimates because of nonhomogeneous conditions (both spatial and temporal variability) within a drainage area. To counteract this problem, the basin was subdivided into 20 homogeneous subunits with similar environmental factors such as precipitation and soil moisture storage. The Thornthwaite equation for evapotranspiration was used because of its simple data requirements and demonstrated success in watershed studies. Further, an equation for snow accumulation and ablation was added. The error between annual calculated runoff and measured runoff was within 5% during six of the calibration years; for one year, the error between calculated and measured runoff increased to 11%. On a monthly basis, calculated runoff exceeded measured streamflow during the period July through October of each year, perhaps due to unaccounted irrigation withdrawals. Calculated runoff underestimated measured discharge during months with heavy and widespread rainfall or rapid snowmelt (typically December through April). Shelton mentioned that an analysis of this type allowed for an assessment of the human influence on runoff; however, he did not conduct such an evaluation for this drainage basin.

Shelton (1989) analyzed the effects of scale differences by comparing calculated runoff for the years 1951 to 1960 over the Deschutes drainage based on the entire basin area, based on nine subbasins (which reduced the root mean square error by 35%), and based on 20 sectors (which reduced the root mean square error by greater than 50%). Modelled runoff improved as a result of dividing the drainage area into smaller, homogeneous areas. For the analysis based on 20 sectors, the greatest difference between monthly calculated runoff and measured runoff was 10 mm over the entire basin. He postulated that overestimation of runoff by the model in the late fall could be due to snow accumulation and ablation in higher elevations. In regard to the accuracy of estimating discharge, he noted that:

> persistence in overestimating or underestimating observed monthly runoff is a common trait when modeling large watersheds, and persistence tends to inflate differences between observed and modeled runoff (Shelton 1989, p. 377).

Owe (1985) investigated the effects of land-use transitions on streamflow in the Chester Creek drainage basin (158 km^2) in Pennsylvania by utilizing the Thornthwaite water balance to predict runoff due to climatic factors and by using aerial photography to monitor land-use changes. In the early 1930's, the predominant land-uses of the watershed were forest, agriculture, and open fields. Over the period 1932 to 1982, he found increases in annual, dormant-season, and growing-season discharge of 51%, 46%, and 57%, respectively. As a result, he assumed that these increases in discharge resulted from increases in urban and suburban land-uses.

Essery and Wilcock (1990) studied a small drainage basin in Northern Ireland and the effects of channel widening, deepening, steepening, and straightening on streamflow. Data from five years before channelization and three years after channelization were used for analyses. Four adjacent subcatchments not affected by channelization were used as controls. The Penman (1948) model of evapotranspiration was utilized, and groundwater measurements were used to assess the accuracy of the water balance. The authors preliminarily determined that streamflow had increased as a result of channelization, yet no evidence of significant increase in low flows was found. The purpose of the channelization was to increase streamflow by lowering water table levels upstream, thereby removing water from floodplain storage; however, groundwater levels in the floodplain were not significantly reduced.

1.3.4 Mississippi River Studies

In the 1950's and 1960's several large scale hydrologic studies of North America were undertaken in order to evaluate the hydroclimatology of the Mississippi drainage basin and its subbasins. For these studies, the Thornthwaite water balance model was used as well as other methods (e.g., the Budyko water balance and an atmospheric vapor flux method). Brief summaries of these studies follow.

Benton, et al. (1950) and Benton and Blackburn (1950) calculated a complete balance of the hydrologic cycle for the Mississippi drainage basin in order to determine the interrelationships of precipitation, evapotranspiration, and runoff. The mean annual precipitation over the watershed was determined from precipitation maps of the U.S. and estimated to be 75.4 cm (28.7 in). The mean annual runoff from the watershed was estimated from stream gauging records to be 16.8 cm (6.6 in). Thus, it was assumed that the mean annual evapotranspiration over the watershed was 58.7 cm (23.1 in). The study estimated that 12% to 14% of the annual precipitation within the watershed was land-derived, and 86% to 88% of the annual precipitation originated from a maritime source.

Benton and Estoque (1954) calculated evapotranspiration over the North American continent for the year 1949 based on integrated transfers of atmospheric water vapor and measured precipitation. The annual evapotranspiration estimated from this method was 56.1 cm (22.1 in). This result was compared to an annual evapotranspiration of 41.9 cm (16.5 in) calculated from the Thornthwaite model. The authors determined that the Thornthwaite evapotranspiration estimates were too low in winter since the method assumed no evapotranspiration occurred whenever a mean monthly temperature was below freezing. Further, the Thornthwaite evapotranspiration produced an early summer maximum while the atmospheric water vapor estimate maximum occurred approximately one month later.

Rasmusson (1967, 1968) conducted a large-scale study of the North American continent and the Central American Sea (the Caribbean Sea and the Gulf of Mexico) utilizing atmospheric water vapor flux data to estimate the water balance of these areas. The vertically integrated distribution of flux divergence was used to calculate mean monthly values of evapotranspiration and soil moisture storage over the period May 1, 1958, to April 30, 1963. These estimates were then compared with values of evapotranspiration calculated with the Thornthwaite (1948) and Budyko (1956) water balance methods. The results indicated that the Thornthwaite method significantly underestimated values of evapotranspiration in winter and overestimated values in summer. Thus, with the Thornthwaite method, soil moisture storage was overly high in winter months and too low during summer months. The Budyko estimates of evapotranspiration were determined to be more consistent with the values calculated from atmospheric vapor flux data although they showed an approximate one-month lag.

Using atmospheric water vapor flux divergence data, Rasmusson (1971) calculated a mean annual rainfall of 60.8 cm, a mean annual streamflow of 9.6 cm, and a mean annual evapotranspiration of 51.2 cm for the Central Plains Region (an area of 4,200,000 km^2 corresponding to the Great Plains). For his Truncated Eastern Region (an area of 1,950,000 km^2 corresponding to the eastern U.S. without including the

Great Lakes), a mean annual precipitation of 107.0 cm, a mean annual streamflow of 44.2 cm, and a mean annual evapotranspiration of 62.8 cm were calculated. For the Ohio Basin (an area of 530,000 km^2), a mean annual precipitation of 112.7 cm, a mean annual streamflow of 45.7 cm, and a mean annual evapotranspiration of 67.1 cm were reported.

Hare (1972) evaluated the annual water balance for the U.S. and Canada based on observed values of precipitation and runoff. Rather than calculating the Thornthwaite potential evapotranspiration, he used observed values of global solar radiation, cloudiness, surface relative humidity, and surface albedo to estimate net radiation and mean annual values of evapotranspiration. A comparison to other evapotranspiration estimates was not presented.

After massive flooding along the Mississippi River in 1973, subsequent hydrological studies emphasized the effects of human modification on the river. For example, Simons, et al. (1974) determined that the channel of the Middle Mississippi had lost approximately one-third of its capacity since 1837 due to channel confinement.

Belt (1975) claimed that regulation of the Mississippi River had produced changes in the river's flow regime. He attributed the 1973 flood on the Middle Mississippi to higher stages which occurred as a result of man-made levees and navigation works which reduced the cross-sectional area of the channel and correspondingly reduced flow-carrying capacity.

Muller (1976) compared water balance estimations of surplus for Lower Mississippi River floods in 1927, 1973, and 1975 in a pilot study to determine whether the flow regime of the Mississippi River had been altered by regulation and land-use changes. In 1927, the Mississippi was, for the most part, unregulated; by 1973, the river was largely controlled by levees and reservoir systems. His study utilized discharge measurements at Vicksburg over three flood seasons (September through April) as compared to a water balance surplus (calculated for climate divisions within the drainage basin). A standard six-inch soil moisture capacity was assumed for all climate divisions within the basin, and a decreasing availability system of soil moisture depletion was used. Drainage throughout the entire basin was based on a two-month lag time. The Middle Mississippi basin generated the most discharge of four subbasins during the 1926-1927 flood season, while the Arkansas basin produced the greatest discharge during the 1972-1973 flood season. The Middle and Lower Ohio basin generated the greatest contribution of discharge during the flood season of 1974-1975. He found that the quantity of modelled runoff for the 1972-1973 flood season was greater than the quantity estimated for the 1926-1927 flood season, despite the fact that maximum stage readings occurred during the 1926-1927 flood.

Kesel (1988) studied changes in the sediment load of the river from 1930 to 1982 and concluded that the suspended sediment discharge of the Mississippi River has decreased by as much as 70 percent since 1850. He listed the developments that have affected sediment quantities in the river as follows: earthquakes; enlargement or closure of distributaries; land-use changes; channel dredging; sand and gravel mining; and construction of dams, levees, revetments, dikes, and cutoffs. A significant decrease in suspended sediment load occurred after the construction of dams in the Missouri and Arkansas basins.

Koellner (1988) considered the effects of climate variability and change on the Mississippi River's navigation system and emphasized that most of the actions that have been taken to develop the Mississippi and Illinois rivers have occurred as a result of climate stresses. In light of this he remarked,

> Changes in such disparate activities as urbanization, channelization of streams, flood control, upland conservation, and agricultural land-use practices will only heighten the importance of regional problems related to climate change and to climate variability (Koellner 1988, p. 275).

Grubaugh and Anderson (1989) examined the effect of a navigation dam on discharge and water-surface elevation in Pool 19, upstream of the Ohio River confluence on the Upper Mississippi River. Data for a 108-year period (1878 to 1986) were analyzed by linear regression for various time intervals. They concluded that increased sedimentation over the 74-year post-dam period resulted in a loss of storage capacity, which in turn, resulted in increased flood days and increased recurrence of major floods.

Gleick (1990) calculated indices of reservoir storage, demand, hydroelectric generation, groundwater use, and streamflow variability in order to speculate about the vulnerability of the Mississippi drainage basin to climatic change. Although Gleick's analysis was based on climate change, changes in water availability due to land-use changes or river regulation could produce similar effects within the basin. Based on data from the U.S. Water Resources Council (1978b) and the U.S. Geological Survey (1986), he concluded that:

1. The Tennessee, Ohio, Arkansas-White-Red, Upper Mississippi, and Lower Mississippi basins have relatively small storage volumes compared to the quantity of annual flow; the Missouri basin has a large storage capacity compared to annual flow;
2. Demand is a large fraction of annual flow in the Missouri basin; demand is only a small fraction of total available water in the Tennessee, Ohio, Arkansas-White-Red, Upper Mississippi, and Lower Mississippi basins;
3. Both the Missouri and the Tennessee watersheds rely heavily on hydroelectricity;
4. Both the Missouri and the Arkansas-White-Red basins rely heavily on groundwater to supplement surface water;
5. The Lower Mississippi, the Arkansas-White-Red, and the Missouri have highly variable streamflows; and
6. The Missouri basin is quite sensitive to climate variability and change.

Kuhl and Miller (1992) used the global climate model developed by the Goddard Institute for Space Studies (GISS) to calculate monthly runoff for 16 of the world's largest rivers, including the Mississippi. The monthly runoff calculated by the model was too high in spring and summer and too low in late summer, fall, and winter as compared to observed values of runoff for the Mississippi basin. Further, the precipitation calculated by the model was too low in the eastern portion of the watershed and too high in the west. Overall, however, the process of averaging quantities over the entire basin tended to cancel these errors such that the mean annual precipitation and runoff calculated by the model matched observed values within 10%.

More researchers are recognizing the need for accurate regional and continental scale water balance models. Dolph and Marks (1992) established a database of precipitation and streamflow for the U.S. using a geographic information system (GIS). Monthly streamflow data for the period 1948 to 1988 from 1,014 unregulated or minimally regulated gauging stations were converted to mean annual depths at each gauge. Drainage basin sizes varied from 4 km^2 to 35,224 km^2. Monthly precipitation measurements for 1,211 stations as well as 405 snow water equivalence gauges from the SCS Snotel database were converted to values of mean annual precipitation. These data were input into a raster-based GIS in order to map the spatial extent and variability of runoff and precipitation across the U.S. They noticed a significant increase (approximately 40%) in measured precipitation by incorporating the Snotel data in the Pacific Northwest, Great Basin, Upper Colorado, and Lower Colorado regions. The mean annual discharge for these regions was 50.6 cm, while the mean annual precipitation (based on precipitation gauges only) was 46.5 cm. By comparison, the mean annual precipitation calculated from both the precipitation and Snotel gauges was 68 cm.

1.4 Global Energy and Water Cycle Experiment

The Global Energy and Water Cycle Experiment (GEWEX) is a current project established by the World Climate Research Programme (WRCP). The hydrologic cycle is a key component in the global energy balance and for that reason is a major factor in evaluating climate change. Global climate models (GCMs) utilize 100 km grid-scale elements, and one of the goals of GEWEX is to develop a hydrological model that is appropriate for use at this scale (Chahine 1992).

The GEWEX program encompasses several objectives:

1. To measure hydrological and energy fluxes via atmospheric and surface properties;
2. To model the global hydrological cycle;
3. To predict the variability of global as well as regional hydrological processes and water resources in regard to environmental change; and
4. To develop data systems which will enable long-range forecasts of weather and climate conditions and effects on water resources.

Initially, the GEWEX Continental-Scale International Project (GCIP) proposed to accurately model the streamflow of the Mississippi River by utilizing a dense network of both atmospheric and hydrologic data. The Mississippi River was determined to be suitable for the study because of a significant rain- and stream-gauge coverage, advanced Doppler weather radars, and continuous wind profilers (WCRP 1993a). A major portion of the study planned to involve data collection and management of the following parameters: air temperature; pressure; wind speed; wind direction; cloudiness; solar radiation; net longwave radiation; evaporation; humidity; precipitation; snow cover; snow water equivalent; soil moisture; groundwater level; streamflow; reservoir storage in major reservoirs; river basin diversions; and consumptive water use.

Currently, the GEWEX project has been downsized to model the Little Washita watershed located within the Arkansas-Red drainage basin. An integrated systems test will be performed on a small scale area before transferring the results to a large scale area (WCRP 1993b).

1.5 Chapter Summary

Water resources management is based on the assumption that the statistical parameters of hydrologic time series (e.g., mean, variance, coefficient of skewness, and autocorrelation coefficients) do not change over time; however, it is important to realize that climate changes, land-use changes, and river regulation influences do have an effect on streamflow. Land-use changes and river regulation have been significant in the Mississippi River basin over the last 60 years or so. Therefore, it is crucial that we better understand the Mississippi River system as a whole and the effects of human-induced modifications upon the river's flow regime.

Comprehensive studies such as GEWEX are of great scientific and societal importance; yet, at the same time, such detailed studies take many years to organize, complete, and evaluate. In contrast, this study requires only measurements of temperature, precipitation, and river discharge and utilizes a water balance methodology to model the Mississippi River basin. Gleick (1989) reviewed studies of the impacts of climate change on water resources and noted the following:

> The World Meteorological Organization (1975) studied a series of simulation models in order to evaluate their strength and weaknesses. They concluded that many models perform well for humid basins but that explicit accounting models such as water balance models were distinctly superior in semiarid and arid basins. Perhaps even more significant was the observation that simpler models showed better results

than complex models when the quality of input data is poor (Gleick 1989, p. 336).

Using the water balance model and available discharge data, an annual water balance residual (actual discharge minus modelled runoff) is calculated in order to separate the effects of river regulation and land-use changes from climate variability. Analysis of the water balance residuals over time should provide insight to the integrated effects of river management and land-use change on discharge within the Mississippi drainage basin.

Chapter 1 has contained a brief introduction and the questions and objectives which define the context of this study. In addition, a literature review has been presented which summarizes representative research efforts within the areas of streamflow variability, water balance analyses, and Mississippi River studies. Chapter 2 defines the methodology used in this study. The Thornthwaite water balance method and calculations are discussed, and the Box-Jenkins statistical methodology for time series analysis is described. Chapter 3 presents the results of these statistical analyses for the Mississippi basin, while Chapter 4 describes the results obtained for the Missouri and Ohio-Tennessee subbasins. A summary and discussion of the results, conclusions, and list of future research topics are contained in Chapter 5.

CHAPTER 2

METHODOLOGY

2.0 Chapter Objectives

The major objective of this chapter is to describe the development and analysis of working data sets for annual precipitation, surplus, modelled runoff, discharge, and the water balance residuals. Brief descriptions of the drainage basins, locations of the gauging stations, and a description of the discharge data used in this study are presented. The Thornthwaite water balance methodology and the climatic data used for calculating the water balance indices of surplus and modelled runoff are described. In addition, this chapter contains a summary of the statistical techniques of linear regression analysis and Box-Jenkins time series analysis.

2.1 Basin Descriptions

The Mississippi River drainage basin is subdivided into six major water-resource regions (U.S. Water Resources Council 1978a): the Upper Mississippi region; the Ohio region; the Tennessee region; the Lower Mississippi region; the Arkansas-White-Red region; and the Missouri region. Figure 2.1 shows a map of the drainage basin and the locations of these water-resource regions. For reference, the location of the Little Washita basin to be modelled by the GEWEX project is also shown. Although the focus of this study is the entire Mississippi River basin, two subbasins are analyzed for comparative purposes. Watershed selection is based on the following criteria:

1. Large geographic coverage. Ideally, drainage basins and their corresponding climate division areas should coincide exactly; however,

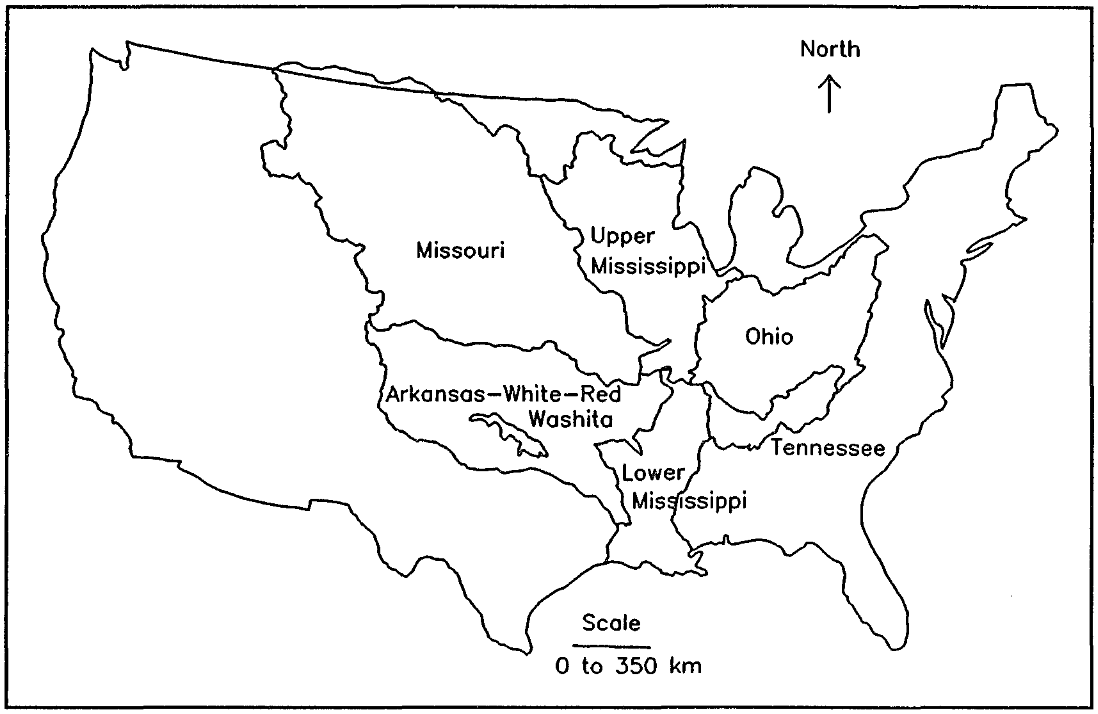

Figure 2.1. Mississippi River drainage basin and subbasins.

climate division boundaries are typically defined by county lines rather than watershed boundaries (except in some western states). Therefore, the selection of larger basins minimizes the relative error in drainage areas due to boundary differences;

2. A continuous streamflow record of adequate length. Daily streamflow data for most gauging stations located on the Mississippi River main stem and its major tributaries are available since January 1928 or earlier from the U.S. Army Corps of Engineers and the U.S. Geological Survey (USGS);

3. Different climatic regimes. The Missouri and the Ohio-Tennessee subbasins are considerably different watersheds in terms of geographic

size, climate, and quantity of streamflow produced. Analysis of these subbasins provides insight into the effects of climatic and human impacts on humid and semiarid watersheds;

4. High quality data. The accuracy of historic streamflow records for the gauges used in this study are considered to be excellent. This means that approximately 95% of the daily mean discharges are reported within a 5% accuracy; and

5. Major human intervention within the watershed area in the form of river management or land-use changes. The natural flow of the Mississippi has been regulated by navigational improvements and flood control programs.

Brief descriptions of each watershed and the corresponding gauging station(s) follow. The locations of selected gauging stations are shown in Figure 2.2.

2.1.1 The Mississippi Basin at Tarbert Landing

The Mississippi River drains approximately 41% of the continental United States. With an overall drainage area of 3,237,500 km^2 (1,250,000 mi^2), the Mississippi basin is the third largest watershed in the world after the Amazon and the Congo river basins (Wells 1980). From 1928 to 1963, the Corps of Engineers maintained a gauge on the Mississippi River main stem at Mile 302.4 (as measured upstream from the Head of Passes at Mile 0.0) near Red River Landing, Louisiana. When the Old River Control Structure (ORCS) was constructed (at Mile 314.6) and put into operation on July 12, 1963, the Lower Old River (at Mile 304) was closed, and

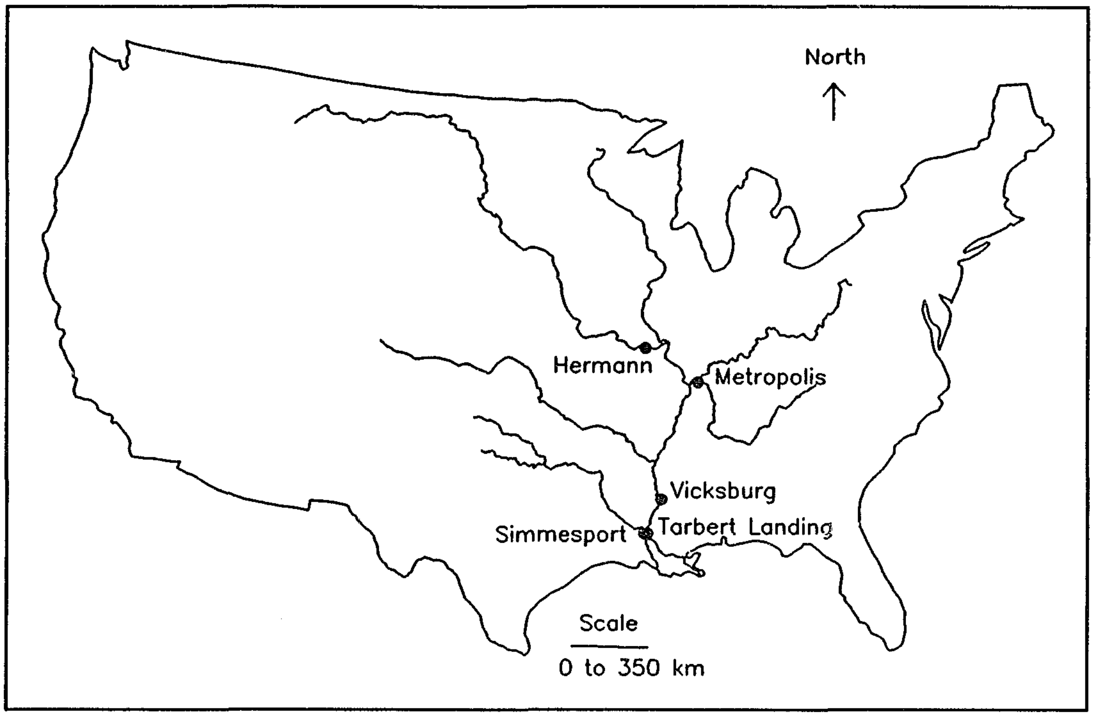

Figure 2.2. Selected gauging stations in the Mississippi drainage basin.

the Red River Landing gauge was relocated slightly upstream (yet still below the diversion) to Tarbert Landing, Mississippi, at Mile 306.3 (USGS Number 07295100). Together, these two gauges form a continuous record of discharge on the Mississippi main stem for the period 1928 to the present. There are no major tributaries below this point on the main stem.

Approximately 25% of the main stem discharge (as measured above the ORCS diversion) flows through the ORCS to the Atchafalaya River (Wells 1980). Since both the Tarbert Landing and Red River Landing gauges are located downstream of the ORCS diversion, discharge from the Atchafalaya River (which includes discharge from the Red River) must be added to the main stem discharge in order to account for the total streamflow from the Mississippi drainage basin. Other researchers have established this precedent of using the Red River Landing/Tarbert Landing gauges as one continuous record and adding streamflow from the Atchafalaya and Red rivers (e.g., Wells 1980; Keown, et al. 1986; Kesel 1988). The Atchafalaya is gauged at Simmesport, Louisiana, (USGS Number 07381490) 4.9 miles as measured downstream from the head of the Atchafalaya River at the confluence of the Red and Lower Old Rivers.

For the years 1932 to 1988, the mean discharge measured at the Tarbert Landing gauge was 13,330 cms (470,600 cfs). The mean discharge measured at the Simmesport gauge was 5,400 cms (190,800 cfs). Hereafter, the sum of discharge from the Tarbert Landing and Simmesport gauges is referred to simply as Tarbert Landing, and the drainage area for the Mississippi River at Tarbert Landing includes the drainage area of the Atchafalaya and Red rivers above Simmesport.

Over the years, diversions have been created to transfer discharge into and out of the watershed for varying reasons, yet these are assumed to be negligible compared to the total discharge from the basin. For example, as directed by the U.S. Supreme Court, water was diverted from Lake Michigan to the Illinois River via the Chicago Sanitary and Ship Canal from January 1, 1900, through 1938 at an average volume of 204 cms (7,200 cfs) in order to flush industrial waste and untreated sewage away from the water supply for the city of Chicago. In 1939, the diversion was reduced to 91 cms (3,200 cfs) (Demissie and Bhowmik 1987).

As another example, the Corps of Engineers has constructed a series of canals to create the Tennessee-Tombigbee waterway. This 377 km (234 mi) waterway connects the Tennessee River at Yellow Creek near Pickwick Lock and Dam to the Black Warrior-Tombigbee Waterway near Demopolis, Alabama. From this point the Tombigbee River continues to the Gulf of Mexico at Mobile. Flow out of the Tennessee River basin into the Tombigbee Waterway is considered to be negligible (Flowers 1990).

2.1.2 The Mississippi Basin at Vicksburg

The Vicksburg gauging station (USGS Number 07289000) is located on the main stem of the Mississippi at Mile 435.7 (as measured upstream from the Head of Passes at Mile 0.0). Two minor tributaries enter the main stem below this point: the Big Black River; and the Homochitto River.

Since the Vicksburg gauge is located above the ORCS diversion, discharges are approximately 20% higher (Wells 1980) than those measured downstream at the Tarbert Landing gauge (without including discharge from the Simmesport gauge). For the years 1932 to 1988, the mean discharge measured at the Vicksburg gauge was 16,590 cms (585,900 cfs).

2.1.3 The Missouri Basin at Hermann

The mouth of the Missouri River is located 195 miles upstream from the mouth of the Ohio River near Cairo, Illinois, on the Mississippi main stem. The Missouri River at Hermann, Missouri, is the largest subbasin of the Mississippi River basin and covers 1,357,678 km^2 (524,200 mi^2). The Hermann gauge (USGS Number 06934500) is located at Mile 97.9 as measured upstream from the mouth of the Missouri. Despite the large area of this subbasin, the discharge contributed to the Mississippi River is small because of a semiarid climate. For the years 1932 to 1988, the mean discharge measured at the Hermann gauge was only 2,210 cms (78,100 cfs). A series of flood control dams were constructed in this subbasin during the years 1953 to 1967, so the degree of regulation within the basin is appreciable (U.S. Geological Survey 1986).

2.1.4 The Ohio-Tennessee Basin at Metropolis

The confluence of the Ohio and the Mississippi rivers occurs near Cairo, Illinois. The Ohio River at Metropolis, Illinois, is the second largest subbasin of the Mississippi basin with a drainage area of 525,770 km^2 (203,000 mi^2) that extends over fourteen states. The Metropolis gauge (USGS Number 03611500) is located at Mile 37.4 as measured from the mouth of the Ohio River. Due to a more humid climate, the Ohio River has the largest discharge of any tributary to the Mississippi River. For the years 1932 to 1988, the mean discharge measured at the Metropolis gauge was 7,610 cms (268,600 cfs).

The Tennessee River is the largest tributary to the Ohio River, entering at Paducah, Kentucky, upstream of Metropolis. The Tennessee Valley Authority (TVA) has made substantial changes within the Tennessee River basin since its creation in 1933. The degree of regulation within the Ohio basin is classified as moderate, while the degree of regulation within the Tennessee basin is considered to be appreciable (U.S. Geological Survey 1986).

2.2 Discharge Data

For this study, mean daily discharge data (in cfs) were obtained directly from the Corps of Engineers Vicksburg District Office for the five gauging stations described above. The Vicksburg District Office also supplied a BASIC computer program (Flowers 1990) to convert mean daily discharge to mean monthly discharge. Additional computer programs written in FORTRAN (Hoff 1990) were used to convert mean monthly discharge for each gauge to annual (water year) depths of water over each drainage basin. All annual quantities used in this study are based on water years. For example, October 1, 1931, through September 30, 1932, represents the 1932 water year.

2.3 Climate Data

The time period of this study is defined by the availability of consistent climate division data. Monthly climate division data for January 1931 to the present are available from the National Climatic Data Center (NCDC) in Asheville, North Carolina. Monthly divisional temperature (in °F) and divisional precipitation (in inches) from January 1931 through December 1988 were obtained on magnetic tape and via telephone connection to the NCDC computer library. The temperature data were corrected by NCDC as described in Karl, et al. (1986) and Karl and Williams (1987). This correction systematically removed data discontinuities resulting from nonclimate effects, e.g., changes in observation times, changes in instrument locations, station relocations, etc., in order to produce a more consistent climate record. The methodology for this correction involved the statistical comparison of stations with discontinuities to nearby homogeneous stations in order to determine whether the discontinuity should be adjusted.

An earlier climate division data set has been developed for the years 1895 to 1930 by the utilization of regression techniques (Karl, et al. 1982). However, the use of regression procedures removes natural variability and introduces some concern in regard to the accuracy of the regression-generated data. Thus, the 1895 to 1930 data set was determined to be inappropriate for inclusion in this study.

2.4 Water Balance Calculations

The water balance was developed in the 1940's by Thornthwaite (1944, 1948) and was further refined by Thornthwaite and Mather (1955, 1957). This approach is a simple accounting of water based on input (from precipitation), outflow (from evapotranspiration and streamflow), and storage (soil moisture). Originally, this method was used for the development of a climate classification system based on an energy and moisture balance; however, other applications are more common at this time, particularly the estimation of streamflow (e.g., Mather 1981; Calvo 1986).

Potential evapotranspiration (PE) represents the climatic demand for water from the environment and is defined as the quantity of water that will evaporate and transpire from a landscape covered by a stand of homogeneous vegetation with no shortage of soil moisture (Muller and Thompson 1987). The Thornthwaite (1948) empirical equation for unadjusted potential evapotranspiration (in cm/month) is

$$e = 1.6\left(10\,\frac{T}{I}\right)^{a},$$

where T is the monthly temperature (in °C), I is an annual heat index created from the sum of 12 monthly heat index values, and a is an empirical quantity calculated according to the equation

$$a = 6.75E{-}07\, I^3 - 7.71E{-}05\, I^2 + 1.792E{-}02\, I + 4.9239E{-}01 \,.$$

The monthly heat index values, i, are calculated as follows,

$$i = \left(\frac{t}{5}\right)^{1.514} ,$$

where t is the mean monthly temperature (in °C). For mean monthly temperatures greater than 26.5°C, the unadjusted potential evapotranspiration is calculated according to the equation

$$PE = -0.043\, t^2 + 3.224\, t - 41.585$$

(Thornthwaite 1948). Mean monthly temperatures equal to or less than 0°C are assumed to result in a potential evapotranspiration of zero for that month.

This unadjusted potential evapotranspiration is based on a standard 30-day month and a standard 12-hour daylength and should be corrected to produce a potential evapotranspiration based on the actual hours per day and days per month for a specific location. A daylength adjustment factor based on the latitude of each division (using a methodology developed by Sellers 1965) is used to adjust potential evapotranspiration by utilizing solar declination to calculate hours per day.

The Thornthwaite potential evapotranspiration produces reasonable estimates on an annual basis, but tends to underestimate potential evapotranspiration during winter and overestimate potential evapotranspiration during summer (e.g., McCabe 1989). However, the advantage of using the Thornthwaite model rather than other models (e.g., Penman 1948; Jensen and Haise 1963) is that minimal and readily available data are required (mean monthly temperature). More accurate models require complex measurements that are typically unavailable over large geographic areas.

Precipitation (P) is the quantity of water available in the environment for evapotranspiration, soil moisture recharge, streamflow, and groundwater recharge. Actual evapotranspiration (AE) is defined as the quantity of water actually used by vegetation within the environment. For this study, soil moisture is utilized on a decreasing availability basis, i.e., the ratio of actual to potential evapotranspiration decreases as a linear function of the quantity of available soil moisture.

During periods when precipitation and soil moisture are not sufficient to meet the potential evapotranspiration demand, a deficit (D) in the water supply to the landscape occurs. Therefore, the moisture deficit represents the quantity of water that would have been used for evapotranspiration by vegetation if sufficient water had been available. Accordingly, the energy demand of the water balance is represented by the equation

$$PE = AE + D .$$

During periods when precipitation and soil moisture are more than sufficient to meet the potential evapotranspiration demand, a surplus (S) in the water supply to the landscape occurs. Therefore, the water balance surplus represents precipitation that is available for streamflow or groundwater recharge. The moisture demand of the water balance is represented by the equation

$$P = AE + S \pm \Delta ST ,$$

where ΔST represents change in soil moisture storage.

The calculation of runoff (RO) accounts for the delay or lag time required for surplus water to infiltrate soil, travel through groundwater supplies, and eventually become streamflow. The lag factor depends on basin size, vegetation cover, soil type, degree of slope, and subsurface soil and rock characteristics (Mather 1981). The standard Thornthwaite and Mather (1955) runoff calculation for large watersheds assumes that 50% of the calculated surplus during a month actually becomes runoff during that month. The remainder is held over and added to the surplus calculated for the next month. Accordingly, half of the next month's surplus (including the amount held over from the previous month) is then allowed to become runoff in the second month and so on.

The time-of-travel through the entire Mississippi basin is approximately two months; however, for consistency in this study, a one-month lag time is used for all basins since a one-month lag is assumed to be the approximate time-of-travel through the Missouri and Ohio-Tennessee subbasins. Modifying the lag factor for a specific basin can result in better agreement between measured discharge and the modelled runoff calculated by the water balance model. It should be noted that the water balance calculations do not take into account snow accumulation or snowmelt. Rather, surplus is assumed to be available for runoff during the same month that the surplus occurs regardless of snowpack conditions.

A standard Thornthwaite monthly water balance computer program in FORTRAN written by Grymes (1988) and based on an earlier program by Willmott (1977) was used to perform the water balance calculations. This program requires inputs of monthly precipitation, monthly temperature, latitude, Thornthwaite heat index, soil moisture storage capacity, and initial values of soil moisture storage and runoff for each climate division.

The soil moisture storage capacities used in this study are the same soil moisture capacities used by NCDC for the calculation of the Palmer Drought Severity Index (Palmer 1965) as reported in the *Weekly Weather and Crop Bulletin* (NCDC 1994). As shown in Figure 2.3, the soil moisture storage capacities within the overall drainage basin range from 12.70 cm (5 in) to 27.94 cm (11 in). Initial values of soil moisture storage and runoff for each division were estimated from 1930 divisional precipitation data (Muller 1988).

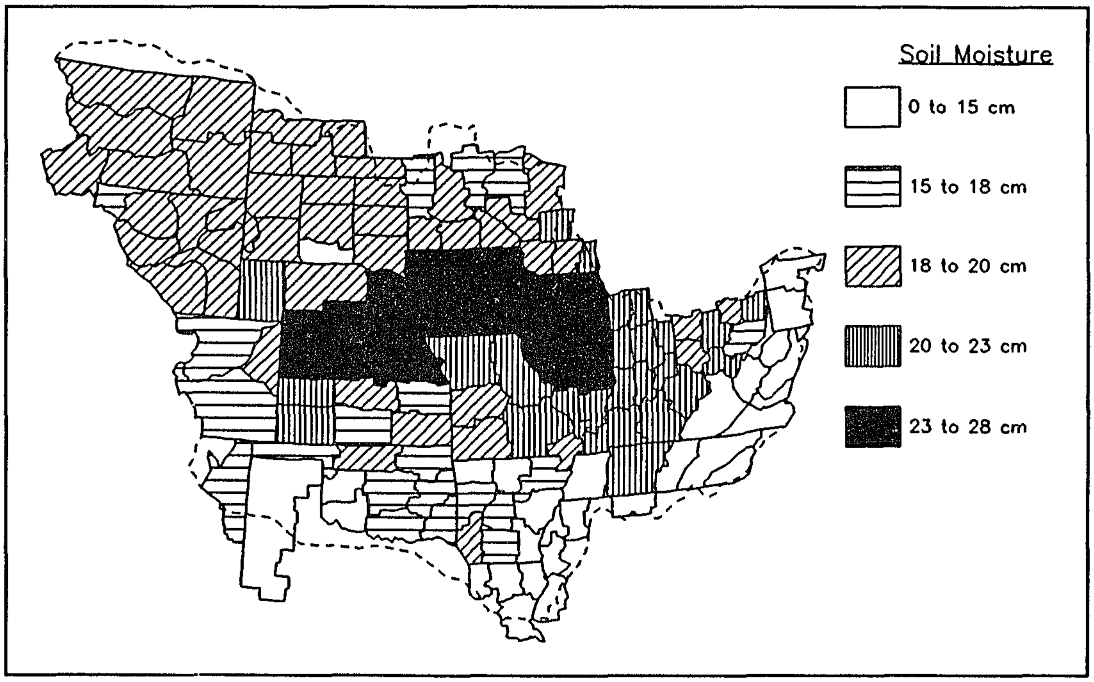

Figure 2.3. Soil moisture storage capacities for climate divisions in the Mississippi basin (NCDC 1988).

The water balance program produces monthly values of potential evapotranspiration, actual evapotranspiration, surplus, and deficit for each climate division. FORTRAN computer programs were designed (Hoff 1990) to calculate monthly runoff (based on a one-month lag) and to convert the monthly values of precipitation, surplus, and modelled runoff to annual (water year) quantities for each climate division. Additional FORTRAN programs were written to convert annual quantities of precipitation, surplus, and modelled runoff for each climate division (weighted by area) to annual depths of water over each watershed.

Composite climate division areas were selected to represent each drainage basin area with the smallest possible error. Table 2.1 contains summary data for each of the drainage basins included in this study. The percent error between watershed area and composite climate division area is calculated for each basin. The composite climate division area for the Mississippi basin agrees extremely well with the area of the drainage basin above the Vicksburg gauge. The greatest error between composite climate division area and drainage basin area occurs for the Missouri basin.

Table 2.1. Comparison of Drainage Areas and Climate Division Areas for the Mississippi Basin and Subbasins

Gaging Station	Drainage Area km^2	Number of Climate Divisions	Division Area km^2	Error
Tarbert Landing/ Simmesport	2,913,494 226,807	147	3,173,649	+1.06%
Vicksburg	2,896,037	135	2,898,583	+0.09%
Hermann	1,357,678	49	1,326,457	-2.30%
Metropolis	525,770	34	520,878	-0.93%

Generally, the water balance produces the best estimates of streamflow for small, homogeneous watersheds; therefore, scale is an important factor. Even though extremely large basins are modelled in this study, the aggregation of spatial data from areas as small as individual climate divisions results in a more accurately modelled runoff. Figures 2.4, 2.5, 2.6, and 2.7 contain maps of the composite climate divisions for the Mississippi basin at Tarbert Landing, the Mississippi basin at Vicksburg, the Missouri basin at Hermann, and the Ohio-Tennessee basin at Metropolis, respectively.

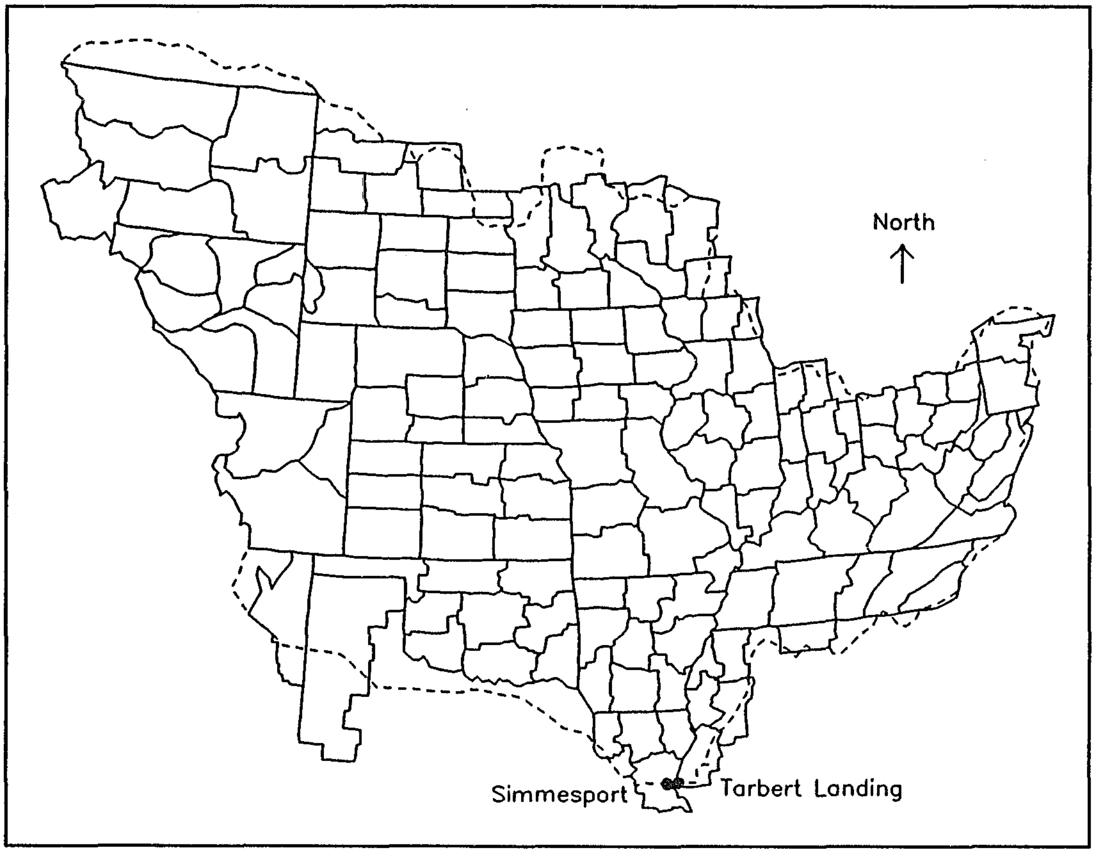

Figure 2.4. Composite climate divisions for the Mississippi basin at Tarbert Landing.

The difference between measured discharge and modelled runoff is calculated as a water balance residual. A positive water balance residual indicates that actual discharge is greater than the modelled runoff estimated for a specific year, i.e., the model underestimates actual streamflow. A negative water balance residual indicates

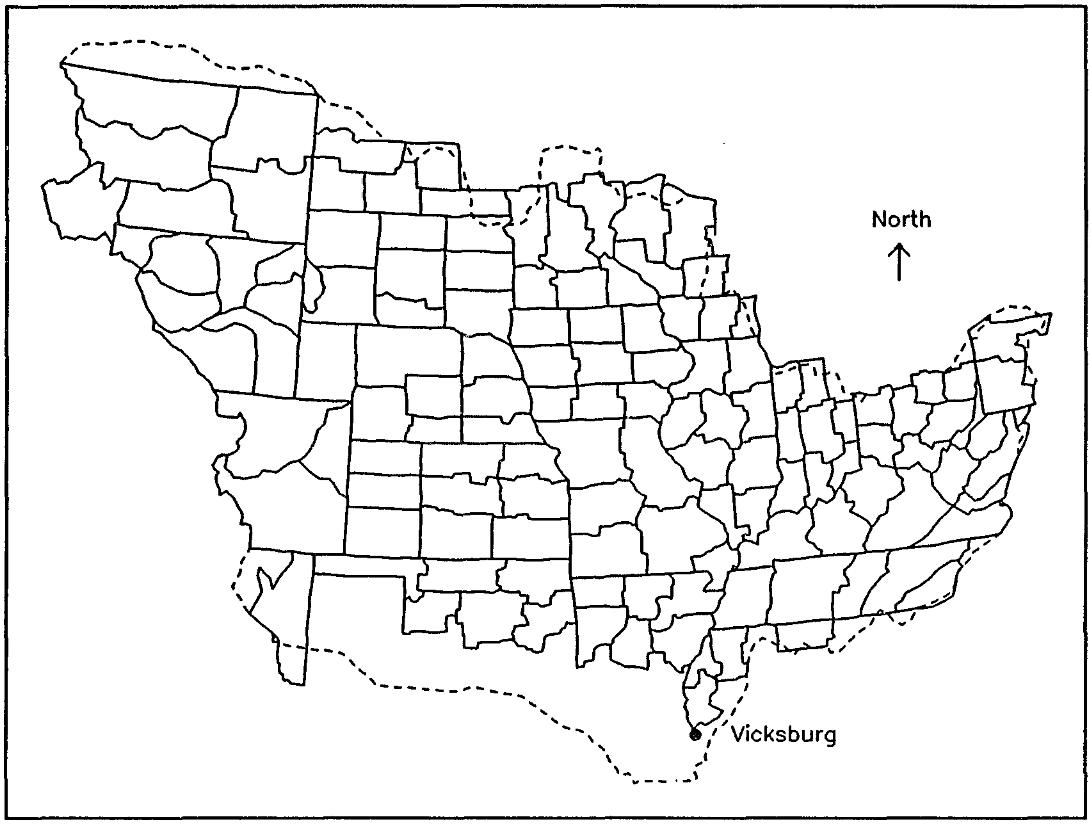

Figure 2.5. Composite climate divisions for the Mississippi basin at Vicksburg.

that actual discharge is less than the modelled runoff estimated for a specific year, i.e., the model overestimates actual streamflow. The water balance residuals represent factors which are not accounted for in the water balance model, including the following (Muller 1976):

1. Errors in discharge measurement (e.g., inaccuracies in the discharge rating curve, inaccuracies in measuring extreme discharge rates, etc.);
2. Errors inherent within the water balance model (e.g., errors generated from inaccuracies in temperature or precipitation measurement, inaccuracies in estimating evapotranspiration, inaccurate lag time, etc.);

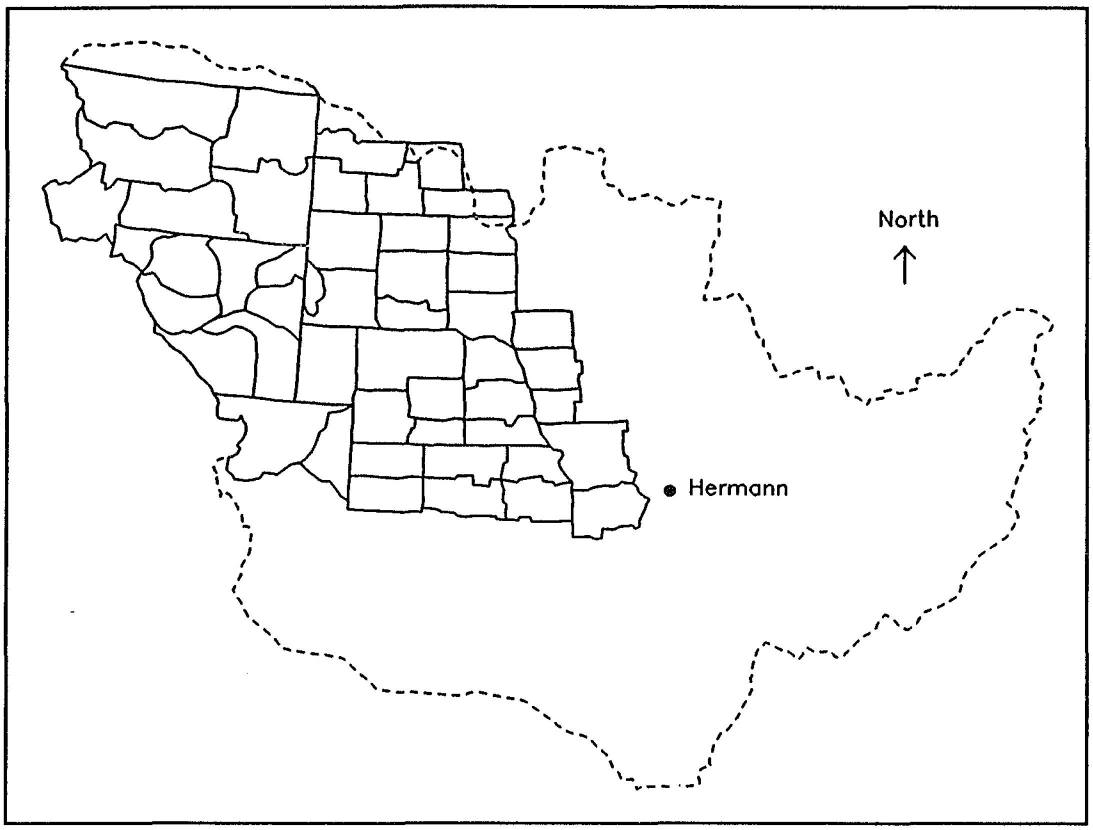

Figure 2.6. Composite climate divisions for the Missouri basin at Hermann.

3. Errors in the overall water balance due to factors which are not accounted for by the model (e.g., groundwater storage and snowpack conditions); and

4. Changes in discharge caused by reservoir storage, river management, and land-use changes.

As mentioned earlier, these factors can occur simultaneously and competitively over time; as such, residual analysis can be difficult.

Cumulative residual mass curves (plots of water balance residuals versus water year) are used to graphically illustrate any changes in the water balance residuals over time. A cumulative residual mass curve reveals minor changes and anomalies that are

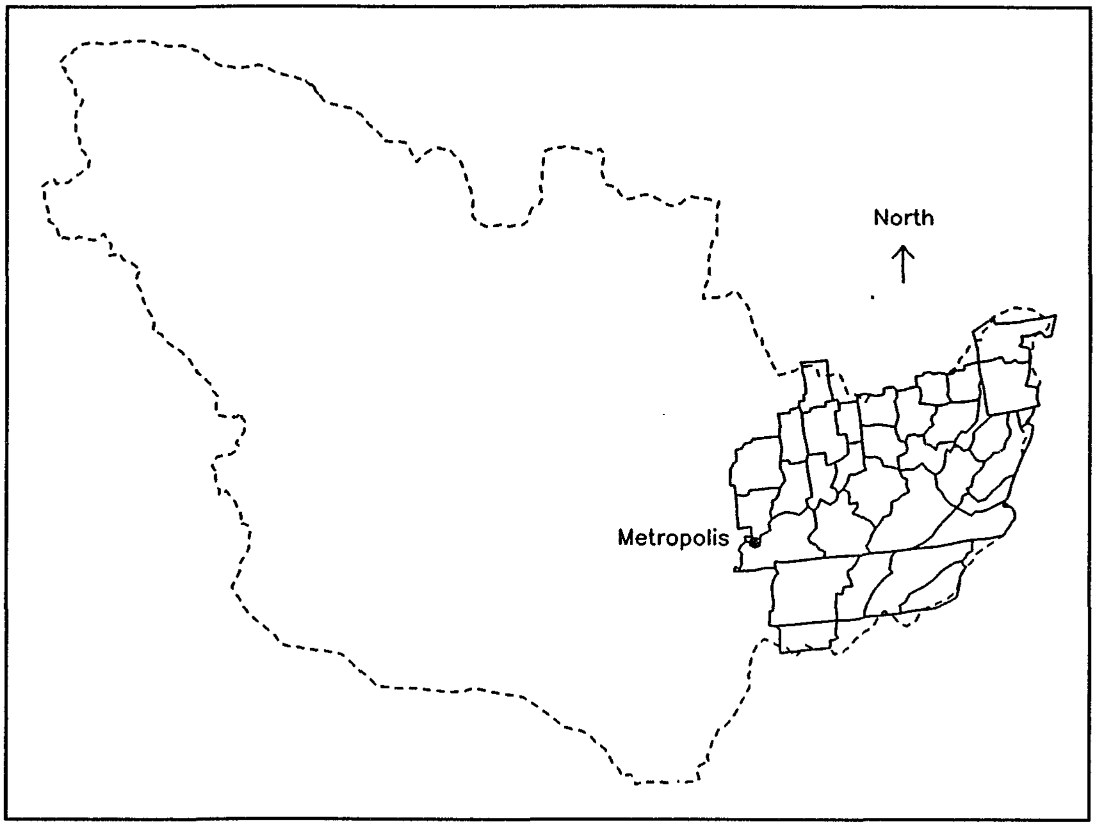

Figure 2.7. Composite climate divisions for the Ohio-Tennessee basin at Metropolis.

sometimes masked in other types of graphs (Owe 1985). If the relationship between measured discharge and modelled runoff remains relatively constant (resulting in residuals of a constant magnitude) over a period of years, the cumulative residuals appear as a straight line. If, however, the relationship changes over time (producing a change in the magnitude of the residuals), the cumulative residuals appear as a curve with maximums and minimums that correspond to changes in the relationship between discharge and modelled runoff.

2.5 Statistical Analysis

The calculations for the statistical analysis of data in this study utilize the Statistical Analysis System (SAS) computer software system, Version 6.07.02. The

SAS procedure PROC UNIVARIATE is used to calculate the means, standard deviations, and other summary statistics for each time series (Schlotzhauer and Littell 1987). In addition, this procedure is used to evaluate the normality of each data set, since a normal population distribution is required for the use of parametric statistical methods. The procedure PROC UNIVARIATE with the NORMAL option performs the Shapiro-Wilk test for normality and calculates the test statistic for normality (W:Normal) as well as its associated probability value (Prob$<$W). Probabilities range from zero to 1 and indicate the relative probability that the sample is drawn from a normal distribution. Probability values less than α indicate that the data are not a sample from a normal population. An α-level of 0.05 is used for evaluating the results of all statistical tests conducted in this study.

Linear regressions are used to compare the water balance indices of surplus and modelled runoff to actual measured discharge to determine the effectiveness of the water balance model in each basin. The equation for a simple linear regression model is

$$y_t = \beta_0 + \beta_1 x_t + \epsilon_t ,$$

where the intercept is β_0, the slope of the straight line is β_1, and ϵ_t is the error term which measures the effect of all factors other than x_t on y_t. The SAS procedure PROC REG performs a least-squares regression which estimates these parameters and other related statistics (Schlotzhauer and Littell 1987). The least-squares point estimates minimize the sum of squares errors (SSE) or

$$SSE = \Sigma \left(y_t - \hat{y}_t\right)^2,$$

where y_t is the dependent variable, and $\hat{y}_t$ is the predicted value of y_t.

An F-test for the overall model tests the null hypothesis that the independent variable x_t does not affect y_t. Thus, an α-level less than 0.05 indicates that the variable x_t significantly affects y_t. SAS calculates both an F value and a probability value for the overall F model. In addition, partial F-statistics are calculated for individual portions of the model.

Regression results are compared by evaluating the root mean square error (MSE) and the adjusted multiple coefficient of determination (adjusted r^2). The root MSE is calculated according to the equation

$$s = \sqrt{\frac{\sum_{t=1}^{n}\left(y_t - \hat{y}_t\right)^2}{n - n_p}}.$$

Further, the adjusted r^2 is calculated as follows:

$$adjusted\ r^2 = \left(\frac{\sum_{t=1}^{n}\left(\hat{y}_t - \bar{y}\right)^2}{\sum_{t=1}^{n}\left(y_t - \bar{y}\right)^2} - \frac{n_p - 1}{n - 1}\right)\left(\frac{n - 1}{n - n_p}\right),$$

where n is the number of observations, and n_p is the number of parameters. The best regression model is determined by the smallest root MSE value and the highest adjusted r^2, which increases only when an important independent variable is added.

To employ the classical regression model, the following assumptions must be met for the regression errors, ϵ_t (Steel and Torrie 1980):

1. Regression errors must be normally distributed;
2. Regression errors must have a constant variance; and
3. Regression errors must be independent.

These assumptions are validated as follows. First, the assumption of normality for the regression errors is tested with SAS by using the procedure PROC UNIVARIATE. Secondly, the assumption of constant variance for the regression errors is tested by plotting the linear regression residuals against x_t, $\hat{y}_t$, and t. For regression errors that have a constant variance, these plots have the appearance of a horizontal band. Thirdly, the assumption of independence for the regression errors is examined by utilizing the Durbin-Watson test for first order autocorrelation. The Durbin-Watson test is performed the procedure PROC REG with the model option DW. First order autocorrelation is more common than autocorrelation occurring at longer lags, so this test is fairly effective. For $\alpha = 0.05$, $k = 1$ (number of independent variables), and $n = 57$ (number of observations), the Durbin-Watson critical value is between $d_L = 1.536$ and $d_U = 1.607$ (Shaw and Wheeler 1985). This establishes the following acceptance and rejection regions for the null hypothesis of zero autocorrelation:

1. 0 to $d_L = 1.536$, rejection region;
2. $d_L = 1.536$ to $d_U = 1.607$, test indeterminate;
3. $d_U = 1.607$ to $(4 - d_U) = 2.393$, acceptance region;
4. $(4 - d_U) = 2.393$ to $(4 - d_L) = 2.464$, test indeterminate;
5. 2.464 to 4.0, rejection region.

It should be noted that in the practical use of real world data, these assumptions probably do not hold exactly and reasonable results can still be obtained (Bowerman and O'Connell 1987).

2.6 Box-Jenkins Time Series Analysis

Trends in general can be attributed to cumulative natural or human-related causes, while periodicities usually can be associated with astronomical cycles, e.g., seasonality (Kite 1989). Cluis, et al. (1989) surveyed the classical statistical methods used for trend detection and noted that most trend detection tests are based on the assumptions that data are normally distributed and temporally independent; however, autocorrelation between time series observations is quite common. In this study the Box-Jenkins univariate approach to time series analysis is utilized to model the annual series developed for each drainage basin since this method accounts for autocorrelation between time series observations.

A time series is a chronological sequence of observations that are most often equally spaced in time. Univariate time series analysis is used to develop a time series model based solely on the past values of a single variable. Forecasting is the prediction of future events based on past information with the assumption that the pattern of past events will persist in the future. Primarily, statistical time series methods have been developed and used for forecasting purposes, yet these methods also are useful for identifying historical patterns in time series data.

Jenkins (1979) advocated the use of time series models (particularly univariate stochastic models of dynamic systems) for the following reasons:

1. Models provide insight into the behavior of the system;
2. Models provide additional understanding in regard to the basic mechanisms influencing the output of the system;
3. Models provide forecasting results;
4. Models provide information that can be used to optimize the future performance of the system;
5. Univariate models may be used to screen data during the early stages of analysis; and
6. Univariate models may provide the only practical approach to time series analysis due to the sheer complexity and magnitude of the system.

For these reasons and particularly because of the complexity of the factors which generate and affect runoff within the Mississippi drainage basin, a univariate time series methodology is utilized in this study.

Box-Jenkins models are classified as either autoregressive, moving average, or mixed autoregressive-moving average. As a group, these models are called autoregressive integrated moving average (ARIMA) models. Autoregression indicates the tendency for the magnitude of a current event to depend on the magnitude of a previous event. The general nonseasonal Box-Jenkins autoregressive-moving average model is described by the equation

$$\phi_p(B)z_t = \delta + \theta_q(B)a_t,$$

where

$$\delta = \mu\,\phi_p(B).$$

Here, μ is the true mean of all possible realizations of the time series, $\phi_p(B)$ is the autoregressive operator of order p, $\Theta_q(B)$ is the moving average operator of order q, and a_t is a random shock.

The backshift operator, B, shifts the subscript of a time series observation backward in time by one time unit. Accordingly, B^k is the backshift operator which shifts the subscript of a time series observation backward in time by k time units. Thus, $\phi_p(B)\,z_t$ represents

$$\phi_p(B)\,z_t = z_t - \phi_1 z_{t-1} - \phi_2 z_{t-2} - \ldots - \phi_p z_{t-p},$$

and $\Theta_q(B)\,a_t$ represents

$$\theta_q(B)\,a_t = a_t - \theta_1 a_{t-1} - \theta_2 a_{t-2} - \ldots - \theta_q a_{t-q}.$$

The random shock, a_t, is selected from a normal distribution which has a mean of zero and a variance that is constant over all time. Further, in an adequate Box-Jenkins model, all (or nearly all) of the autocorrelation between observations is accounted for in the model, such that the random shocks a_1, a_2, ..., a_t are assumed to be statistically independent. Therefore, the random shock series is considered to be a white noise process, i.e., erratic, unexplained, and unpredictable movements that follow no recognizable pattern.

SAS econometric time series (SAS/ETS) software is used to analyze time series data and develop univariate Box-Jenkins models for the following annual series calculated for each watershed: precipitation, modelled runoff, discharge, and the water balance residuals. In addition, models are developed for the fraction of total discharge generated by both the Missouri and Ohio-Tennessee subbasins. At least fifty historical observations are required to construct an accurate, nonseasonal Box-Jenkins univariate model (Bowerman and O'Connell 1987), and sufficient data are available in this study.

The Box-Jenkins methodology develops linear stochastic models based on a four step iterative process as follows (Box and Jenkins 1976):

1. Tentative identification of a model based on historical data;
2. Estimation of model parameters for the tentative model;
3. Diagnostic checking and selection of a new model if the tentative model can be improved; and
4. Forecasting of future time series values after a final model is obtained.

The SAS procedure PROC ARIMA is used to calculate model parameters and associated diagnostic statistics as described below (Brocklebank and Dickey 1986).

2.6.1 Model Identification

Since classical Box-Jenkins models are used to describe stationary time series, the first step in developing a Box-Jenkins univariate model is to determine whether a time series is stationary or nonstationary over the years studied. A time series is stationary if its statistical properties (e.g., mean and variance) are constant through time. Accordingly, a nonstationary time series has statistical properties that vary over time. A nonseasonal, nonstationary time series, y_t, usually can be transformed into a stationary working series, z_t, by the process of first differencing,

$$z_t = y_t - y_{t-1} ,$$

where $t = 2, 3, \ldots, n$, or second differencing,

$$z_t = (y_t - y_{t-1}) - (y_{t-1} - y_{t-2}),$$

where t = 3, 4, ..., n. Initially, graphs of time series observations versus water year are visually examined to determine whether the values fluctuate around a constant mean with constant variance. Further, stationarity is determined by the behavior of the sample autocorrelation (SAC) function.

The SAC function is a graph or listing of sample autocorrelations at lags k = 1, 2, 3, ..., n. The sample autocorrelation, r_k, at lag k measures the linear relationship between observations separated by k time units and is calculated according to the equation

$$r_k = \frac{\sum_{t=b}^{n-k} (z_t - \bar{z})(z_{t+k} - \bar{z})}{\sum_{t=b}^{n} (z_t - \bar{z})^2},$$

where

$$\bar{z} = \sum_{t=b}^{n} \frac{z_t}{(n - b + 1)}.$$

The value of r_k ranges from +1 for perfect positive autocorrelation (which indicates that observations separated by lag k are linear with a positive slope) to -1 for perfect negative autocorrelation (which indicates that observations separated by lag k are linear with a negative slope).

A spike is said to exist in the SAC if the sample autocorrelation at a specific lag is statistically large. This is measured by the t_{rk}-statistic, which is defined as

$$t_{r_k} = \frac{r_k}{s_{r_k}},$$

where s_{rk} is the standard error of r_k. The standard error is calculated by the equation

$$s_{r_k} = \frac{\left(1 + 2\sum_{j=1}^{k-1} r_j^2\right)^{1/2}}{(n - b + 1)^{1/2}}.$$

SAS calculates both r_k and s_{rk} for each lag, but does not calculate the t_{rk}-statistic; however, this value is easily calculated by hand.

Generally, for a nonseasonal time series, a spike exists at low lags (lags k = 1, 2, and 3) in the SAC when the absolute value of t_{rk} is greater than 1.6. For high lags (lags k > 3), a spike exists when the absolute value of t_{rk} is greater than 2 (Pankratz 1983). Therefore, the behavior of the SAC is described by the following (for nonseasonal data):

1. The SAC cuts off after lag k when there are no spikes in the SAC at lags greater than k; and
2. The SAC dies down when the function does not cut off and instead decreases either fairly quickly or extremely slowly.

A time series for which the SAC dies down extremely slowly is considered to be nonstationary and must be transformed into a stationary series before a tentative model can be identified. As described earlier, first differencing or second differencing can be used to transform nonseasonal time series values.

The sample partial autocorrelation (SPAC) function is a graph or listing of sample partial autocorrelations at lags k = 1, 2, 3, ..., n. The sample partial autocorrelation, r_{kk}, measures the autocorrelation of observations separated by a lag of k units with the effects of the intervening observations removed. The sample partial autocorrelation is calculated by the equations

$$r_{kk} = r_1 ,$$

where k = 1, and

$$r_{kk} = \frac{r_k - \sum_{j=1}^{k-1} r_{k-1,j}\, r_{k-j}}{1 - \sum_{j=1}^{k-1} r_{k-1,j}\, r_j} ,$$

where k = 2, 3, ..., n, and

$$r_{kj} = r_{k-1,j} - r_{kk}\, r_{k-1,k-j} ,$$

for j = 1, 2, ..., k - 1. The value of r_{kk} ranges from +1 (which indicates that observations separated by lag k with the effects of the intervening observations removed are linear with a positive slope) to -1 (which indicates that observations separated by lag k with the effects of the intervening observations removed are linear with a negative slope).

A spike exists in the SPAC if the sample partial autocorrelation at a specific lag is statistically large. This is determined by the value of the $t_{r_{kk}}$-statistic, which is calculated by the following equation

$$t_{r_{kk}} = \frac{r_{kk}}{s_{r_{kk}}} ,$$

where $s_{r_{kk}}$ is the standard error of r_{kk}. The standard error is calculated from the equation

$$s_{r_{kk}} = \frac{1}{(n - b + 1)^{1/2}}.$$

SAS calculates both r_{kk} and s_{rkk} for each lag, but does not calculate the t_{rkk}-statistic; again, this value is easily calculated by hand.

Generally, for a nonseasonal time series, a spike exists in the SPAC when the absolute value of t_{rkk} is greater than 2 (Pankratz 1983). Therefore, the behavior of the SPAC is described by the following (for nonseasonal data):

1. The SPAC cuts off after lag k when there are no spikes in the SAC at lags greater than k; and
2. The SPAC dies down when the function does not cut off and instead decreases steadily in a damped exponential fashion (with or without oscillation), a damped sine wave fashion, or a combination of these behaviors.

The behavior of the SAC and the SPAC is used to identify a tentative model according to general guidelines (see Table 2.2). A tentative model is estimated by determining whether $\phi_p(B)$ should be included in the model and the order of the operator (p), whether $\Theta_q(B)$ should be included in the model and the order of the operator (q), and finally, whether δ should be included in the model.

The constant δ, the autoregressive operator, and the differencing operator determine the basic nature of the Box-Jenkins model. The moving average operator determines the effects of random shocks on the basic nature of the model. If the stationary working series z_t are the original time series values, including μ in the model indicates that the observations fluctuate around a constant nonzero mean. The absence of μ indicates that the observations fluctuate around a zero mean.

Table 2.2. Guidelines for the Identification of Nonseasonal Box-Jenkins Models (after Bowerman and O'Connell 1987)

Behavior of the SAC and SPAC	Nonseasonal Operators
SAC has spikes at lag q and SPAC dies down	Moving average operator of order q
SAC dies down and SPAC has spikes at lag p	Autoregressive operator of order p
SAC has spikes at lag q and SPAC has spikes at lag p	Either moving average operator of order q or autoregressive operator of order p
SAC has no spikes and SPAC has no spikes	No nonseasonal operator
SAC dies down and SPAC dies down	Moving average operator and autoregressive operator

If the stationary time series values z_t are differences of the original values, then the inclusion of the constant term δ in a Box-Jenkins model indicates that a deterministic trend exists in the data that is presumed to continue into the future. A deterministic trend implies that the time series parameters are not changing over time, and future values of the time series can be exactly predicted based only on past observations. Thus, a deterministic trend indicates a tendency for time series observations to move persistently in a particular direction. If the stationary time series values z_t are differences of the original values, then the absence of the constant term δ indicates that any trend in the data is stochastic. A series is said to be stochastic when the model parameters are changing over time according to the laws of chance.

Therefore, in a stochastic time series, future values have a probability distribution based on past values. Most hydrologic processes are stochastic or a combination of deterministic and stochastic processes; purely deterministic processes can only result under controlled conditions (Yevjevich 1972).

2.6.2 Estimation of Model Parameters

Although Box and Jenkins (1976) advocated the use of a maximum likelihood approach to calculate point estimates, SAS/ETS calculates the least squares point estimates of the model parameters (such that the values of the parameters minimize the SSE). If the random shocks a_t are normally distributed, then the least squares point estimates are close approximates to the maximum likelihood estimates (Bowerman and O'Connell 1987).

A model is parsimonious when it adequately represents the historical data without including any unnecessary parameters. The following guidelines are used to insure the development of a parsimonious model (Bowerman and O'Connell 1987). For an α-level of 0.05, the absolute t-value calculated for any point estimate must be greater than 2 in order to include that parameter in the model. SAS calculates the t-statistic or t-value (T Ratio) for each parameter as follows,

$$t_{\hat{\theta}} = \frac{\hat{\theta}}{s_{\hat{\theta}}},$$

where Θ is any parameter, $\hat{\theta}$ is the point estimate of Θ, and $s_{\hat{\theta}}$ is the standard error of the point estimate $t_{\hat{\theta}}$. Thus, if $t_{\hat{\theta}}$ is greater than 2, then Θ is statistically large and should be included in the Box-Jenkins model. Similarly, the constant term δ should be included in the model only if the absolute t-value calculated for the point estimate of μ is greater than 2. Recall that

$$\delta = \mu \, \phi_p(B).$$

If $\phi_p(B)$ is not included in the model, then $\phi_p(B) = 1$, and $\delta = \mu$. If $\phi_p(B)$ is included in the model, then SAS calculates an estimated value for δ.

A correlation matrix for the estimated model parameters is also calculated by SAS. High correlations between point estimates indicate that the estimated values may be poor approximations. Generally, the correlations between the point estimates of a model should be less than 0.9 to produce the best model.

The overall standard error for a time series model is calculated according to the equation

$$s = \sqrt{\frac{\sum_{t=1}^{n} (y_t - \hat{y}_t)^2}{n - n_p}},$$

where n is the number of observations in the original time series, and n_p is the number of parameters in the model. Typically, the Box-Jenkins model with the smallest overall standard error is the best model.

2.6.3 Diagnostic Checking

Diagnostic checking involves an evaluation of the following criteria for each tentative model: the Ljung-Box statistic and its associated probability values; the residual sample autocorrelation (RSAC) function; the residual sample partial autocorrelation (RSPAC) function; and stationarity and invertibility conditions. The Ljung-Box statistic, Q^*, indicates the amount of autocorrelation among the residuals from the Box-Jenkins models. SAS calculates the value of Q^* for lags 6, 12, 18, and 24 according to the equation

$$Q^* = n'(n' + 2) \sum_{l=1}^{K} (n' - l)^{-1} r_l^2(\hat{a}),$$

where $n' = n - d$, d is the degree of nonseasonal differencing used to transform the original series, and $r_l^2(\hat{a})$ is the sample autocorrelation of residuals separated by a lag of l time unit. Thus, for an adequate model, Q^* is small, which indicates that the autocorrelations of the residuals are small. A large value of Q^* reveals that the Box-Jenkins model is inadequate since the autocorrelations of the residuals are large. SAS also calculates a probability value (Prob) for Q^* which is used to evaluate the adequacy of a tentative model. The larger the probability value, the more adequate the model. A probability value less than α indicates that the model is inadequate and should be replaced.

The RSAC and RSPAC functions are examined in order to verify that a Box-Jenkins model is adequate. If the probability values calculated for Q^* are less than α, spikes in the RSAC and RSPAC reveal information about an improved model. A spike is said to exist in the RSAC if the t-value is greater than 1.25 at low lags (lag k = 1, 2, 3) and greater than 1.6 at high lags (lag k >3). A spike exists in the RSPAC if the t-value is greater than 2. If the probability values for Q^* are greater than α, only an absolute t-value greater than 2 is considered to be a spike in both the RSAC and RSPAC (Pankratz 1983).

The Box-Jenkins methodology also requires that conditions of stationarity and invertibility are satisfied when determining a final model. Stationarity has been discussed earlier as a time series having constant statistical properties over time. Invertibility implies that the weights placed on the operators in a model are larger for more recent observations and smaller for more distant observations. Specifically, there are stationarity conditions on the autoregressive operators such that the sum of the parameters ϕ_1, ϕ_2, ..., is less than 1. Further, there are invertibility conditions on the moving average operators such that the sum of the parameters Θ_1, Θ_2, ..., is less than 1. Additional stationarity and invertibility conditions are examined for each final model.

2.6.4 Forecasting

In general, the constant term and the autoregressive and differencing operators determine the basic nature of the forecast, while the moving average operators determine how random shocks change the basic nature of the forecast. Although forecasting is not the primary focus of this study, future time series values and 95% prediction intervals are calculated for each series for comparative purposes. SAS calculates a forecast value with a standard error as well as a 95% forecast interval for up to four future values. Historical time series observations are compared to the 95% prediction interval as a means to identify unusual events in each series.

2.7 Chapter Summary

This chapter has presented a methodology for calculating annual quantities of precipitation, surplus, modelled runoff, discharge, and water balance residuals for the

Mississippi basin and subbasins to evaluate changes over the years 1932 to 1988. Specifically, the following statistical tests are employed in this analysis:

- Shapiro-Wilk normality test;
- Least-squares linear regression analysis; and
- Box-Jenkins time series analysis.

CHAPTER 3

THE MISSISSIPPI BASIN

3.0 Chapter Objectives

A major objective of this chapter is to describe the hydroclimatology of the overall basin in both spatial and temporal terms. As a result, annual divisional precipitation and surplus are mapped and discussed in order to provide a geographical and climatological framework for interpreting later results.

Another major objective is to evaluate how accurately the water balance model estimates annual streamflow for the Mississippi basin. In this chapter two versions of the overall Mississippi basin are considered -- the drainage area above Tarbert Landing and the drainage area above Vicksburg. The water balance indices of surplus and modelled runoff are compared with actual discharge data in order to evaluate the effectiveness of the water balance model.

In order to determine whether the hydroclimatology of the watershed has changed over time, Box-Jenkins time series models are developed for nonseasonal (water year) time series data. The statistical and physical implications of each model are discussed.

Finally, an additional objective of this chapter is to determine if human modifications have affected discharge from the overall Mississippi basin. Box-Jenkins time series analysis is used to evaluate the residuals from the water balance model to determine whether changes have occurred during the 57-year period of study.

3.1 Water Balance Results for the Mississippi Basin at Tarbert Landing

Figure 3.1 is a map of the Mississippi watershed which shows the mean annual (water year) precipitation for each of the 147 climate divisions included in this study.

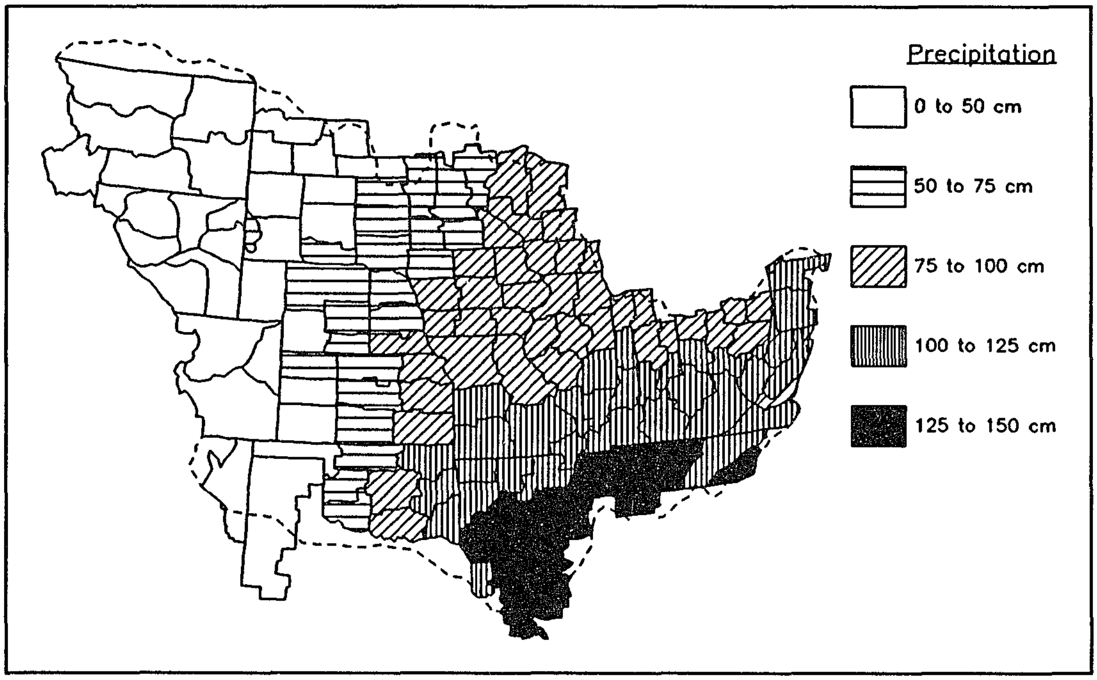

Figure 3.1. Mean annual precipitation for the climate divisions of the Mississippi basin: 1932-1988.

Spatially, mean annual precipitation varies from 149.89 cm for Louisiana Division 5 to 24.62 cm for Wyoming Division 4. As expected, precipitation is larger in the more humid southern and eastern portions of the watershed and smaller in the semiarid western and northwestern reaches of the basin. In fact, the precipitation gradient from east to west is the most notable feature of this map since precipitation drops from greater than 100 cm to approximately one-third of that quantity at the western edge of the Great Plains. The gradient from south to north is also remarkable, since precipitation ranges from nearly 150 cm to less than half of that quantity in the Upper Mississippi Valley.

For comparative purposes, Figure 3.2 illustrates the mean annual water balance surplus calculated for each division. Here, the largest mean annual surplus is generated in the higher altitudes of North Carolina Division 1 (66.87 cm) and Tennessee Division 2 (66.81 cm). Other divisions with large annual surplus values are West Virginia Division 4 (62.34 cm), Alabama Division 1 (60.30 cm), Louisiana Division 5 (59.88 cm), and Mississippi Division 7 (59.73 cm). In stark contrast, several divisions in the western portion of the watershed have a mean annual surplus of zero, i.e., Montana Divisions 3 and 6, and Wyoming Divisions 4, 7, and 9. It should be noted that within the overall Mississippi drainage basin, the majority of surplus water available for streamflow or groundwater recharge is generated within the Ohio-Tennessee watershed. By comparison, very little surplus originates in the Missouri basin. Accordingly, total discharge from the Ohio-Tennessee watershed is much greater than that from the Missouri basin.

To demonstrate the extremes of variability that are possible, Figures 3.3 and 3.4 illustrate the annual water balance surplus for climate divisions within the Mississippi basin for the years 1934 and 1973, respectively. Water year 1934 is an example of a year when severe drought conditions occurred over much of the watershed. In the 1934 map, over half (78 of 147) of the climate divisions had an annual surplus of zero. The remaining climate divisions had annual surplus values ranging from 0.13 cm (in Wisconsin Division 9) to 50.11 cm (in Louisiana Division 2).

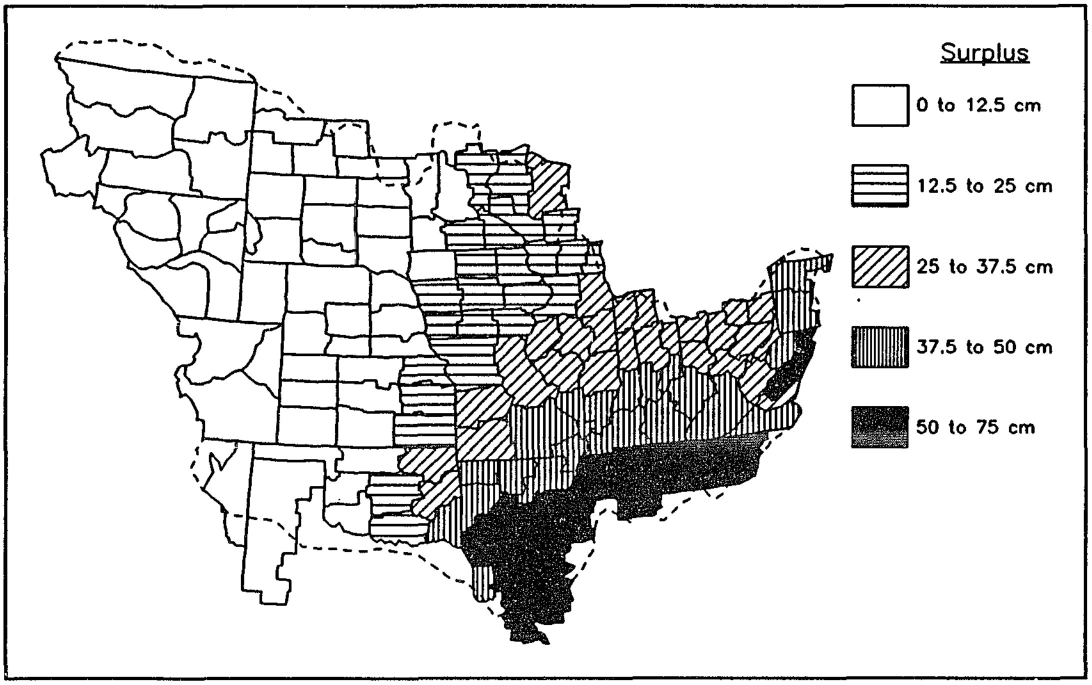

Figure 3.2. Mean annual water balance surplus for the climate divisions of the Mississippi basin: 1932-1988.

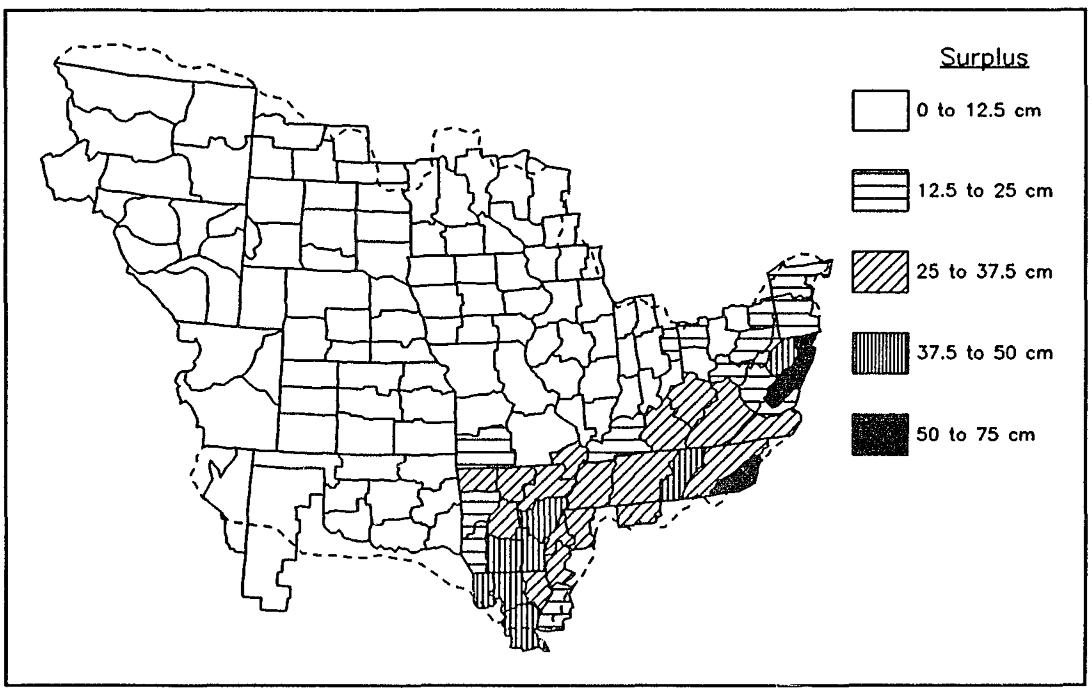

Figure 3.3. Annual water balance surplus for the climate divisions of the Mississippi basin in water year 1934.

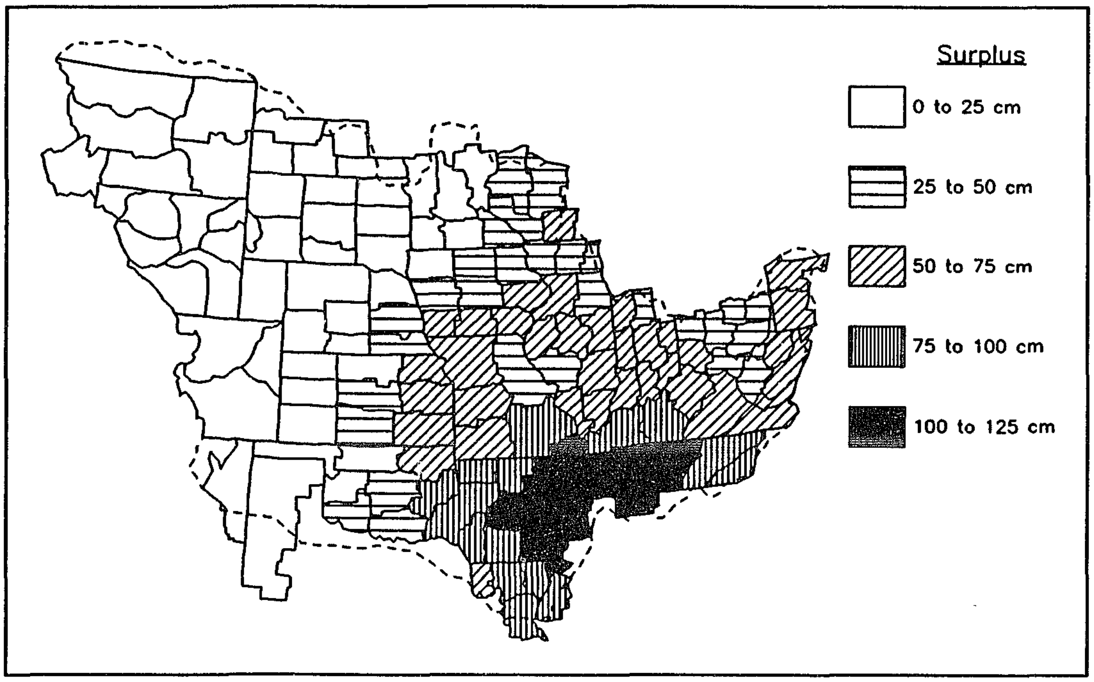

Figure 3.4. Annual water balance surplus for the climate divisions of the Mississippi basin in water year 1973.

In dramatic contrast, water year 1973 is an example of a year in which major flooding occurred in the drainage basin. The Mississippi flood of 1973 caused 28 deaths, forced the evacuation of 50,000 people, and produced estimated damages of over $400 million (Chin, et al. 1975). In the 1973 map, only 28 of the 147 climate divisions had a zero annual surplus. Annual surplus for the other climate divisions ranged up to a maximum of 118.44 cm (Mississippi Division 2). Again, in both of these years, the majority of surplus was generated in the humid eastern and southern areas of the watershed.

In order to examine temporal variability for the watershed as a whole, spatial variability is removed by averaging the quantity of annual precipitation, calculated surplus, modelled runoff, or measured discharge over the area of the basin to produce a mean depth of water. Table 3.1 lists the means and standard deviations for the annual time series as well as the test statistics (W:Normal) and probability values (Prob > W) for normality. At an α-level of 0.05, the hypothesis that these series are samples from normal populations cannot be rejected since the probability values for normality are much greater than 0.05.

Table 3.1. Normality Test Results for the Mississippi Basin at Tarbert Landing Annual Series

Annual Series	Mean cm	Std. Dev. cm	W:Normal	Prob > W
Precipitation	75.46	7.77	0.9743	0.4686
Surplus	19.12	5.42	0.9824	0.7833
Modelled runoff	19.12	5.36	0.9821	0.7723
Discharge	18.82	4.85	0.9795	0.6714
Water balance residuals	-0.30	1.22	0.9715	0.3714

Figure 3.5 contains time series graphs for annual precipitation and surplus expressed as an average depth over the watershed for water years 1932 to 1988. Both of these time series appear to be stationary with relatively constant means and variances. The mean annual precipitation for the Mississippi drainage basin above Tarbert Landing is 75.46 cm (with a standard deviation of 7.77 cm). Temporally, annual precipitation for the entire watershed ranges from a high of 93.63 cm in 1973 (124% of the mean precipitation) to a low of 58.20 cm in 1934 (77% of the mean precipitation). This graph illustrates a high variability in precipitation from year to year, especially during the early years of the series.

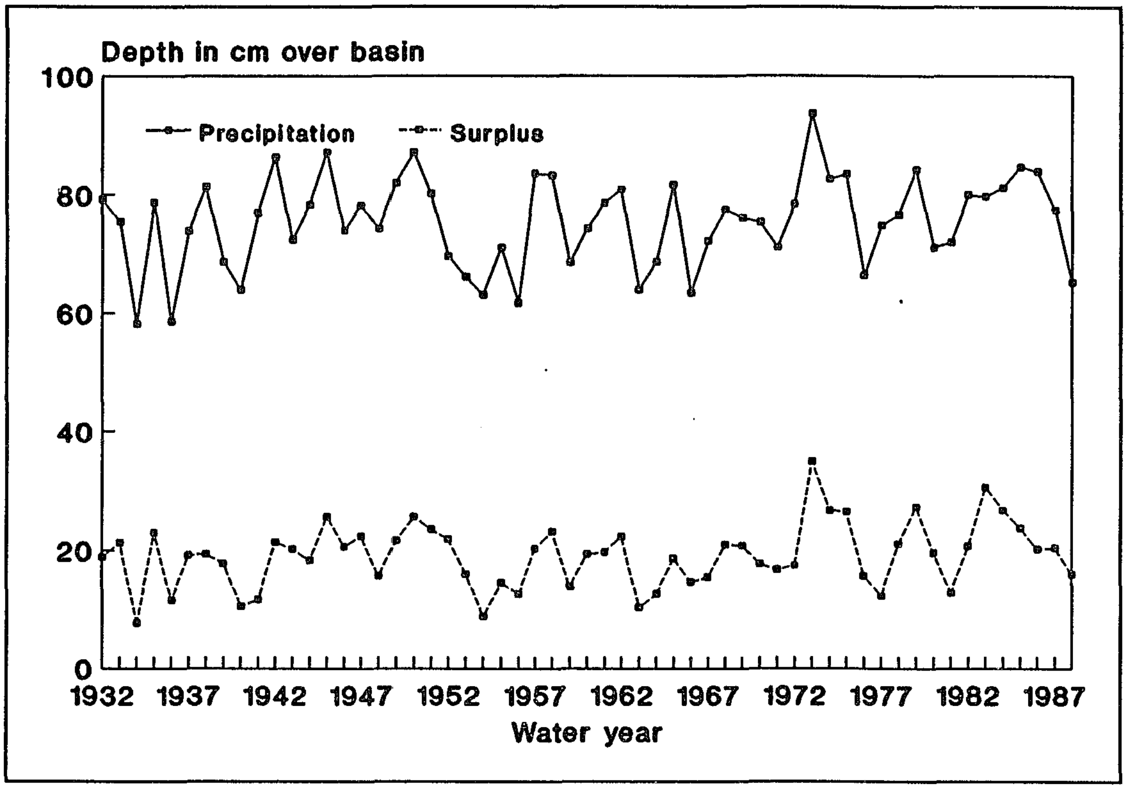

Figure 3.5. Time series graph of annual precipitation and water balance surplus for the Mississippi basin at Tarbert Landing.

The mean annual water balance surplus averages approximately 25% of the mean annual precipitation at 19.12 cm (with a standard deviation of 5.42 cm). The same years that produce maximum and minimum precipitation values also produce maximum and minimum values of surplus at 34.87 cm (46% of the mean precipitation) and 7.75 cm (10% of the mean precipitation), respectively. In more recent years, variability in surplus has been exaggerated by the large quantities generated in 1973, 1974, 1979, 1983, and 1984 as compared to the minimal surplus available in 1977, 1981, and 1988.

As shown in Table 3.1, surplus and modelled runoff have equal annual means; however, in a specific water year, these indices differ slightly as a result of the lagged distribution of runoff. The overall effect of incorporating a one-month lag into the water balance surplus is to reduce variability in the modelled runoff (as shown by a slight 1% reduction in the standard deviation for modelled runoff).

Figure 3.6 compares mean annual discharge to modelled runoff for the Mississippi basin above Tarbert Landing. As illustrated in this graph, modelled runoff estimates streamflow quite well, despite the high variability in annual discharge. The mean annual modelled runoff is nearly 2% larger than mean discharge (with a mean of 19.12 cm) and 11% more variable (with a standard deviation of 5.36 cm). The range of modelled runoff varies from 34.90 cm in 1973 (actual discharge is overestimated by 10%) to 7.90 cm in 1934 (annual discharge is underestimated by 20%). For the years studied, the annual discharge has a mean value of 18.82 cm (with a standard deviation of 4.85 cm) and ranges from a maximum of 31.64 cm in 1973 (168% of the mean discharge) to a minimum of 9.39 cm in 1954 (50% of the mean discharge). Variability in the 1970's and 1980's is a notable feature of the discharge series.

Because the water balance model is so effective at estimating streamflow on an annual basis, the average magnitude of the water balance residuals (calculated from the absolute values of the residuals) is quite small at 0.91 cm as compared to the magnitude of measured discharge (less than 5% of the mean annual discharge). Figure 3.7 is a time series plot of the water balance residuals which shows that the series fluctuates around a mean level of -0.30 cm. The standard deviation is extremely large (1.22 cm), and the range for the series extends from a maximum of 2.69 cm in 1963 (14% of the mean discharge) to a minimum of -4.25 cm in 1947 (23% of the mean discharge).

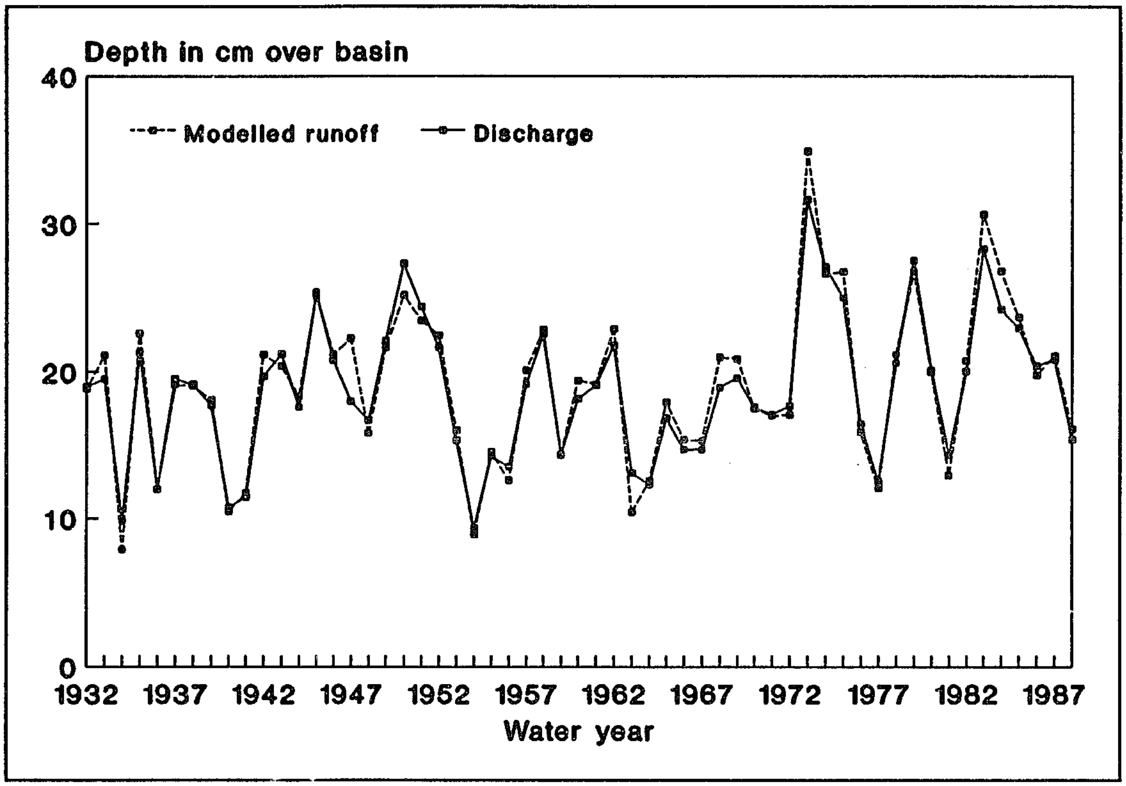

Figure 3.6. Time series graph of annual discharge and modelled runoff for the Mississippi basin at Tarbert Landing.

An examination of the years with large absolute value water balance residuals indicates that possibly the model is not as reliable for estimating runoff during severe climatic conditions. For example, in the flood year of 1973, annual discharge was the highest of the years studied with a total depth of 31.64 cm over the basin. The water balance calculates a modelled runoff of 34.90 cm for the year, which corresponds to an overestimation of 3.26 cm (an overestimation of approximately 10%). It is possible that during 1973 the Corps of Engineers increased reservoir levels throughout the basin to lessen flooding impacts along the Lower Mississippi main stem, and this action would have resulted in a lower total discharge. Or, the additional surplus estimated by the water balance model could have been used for groundwater recharge within the

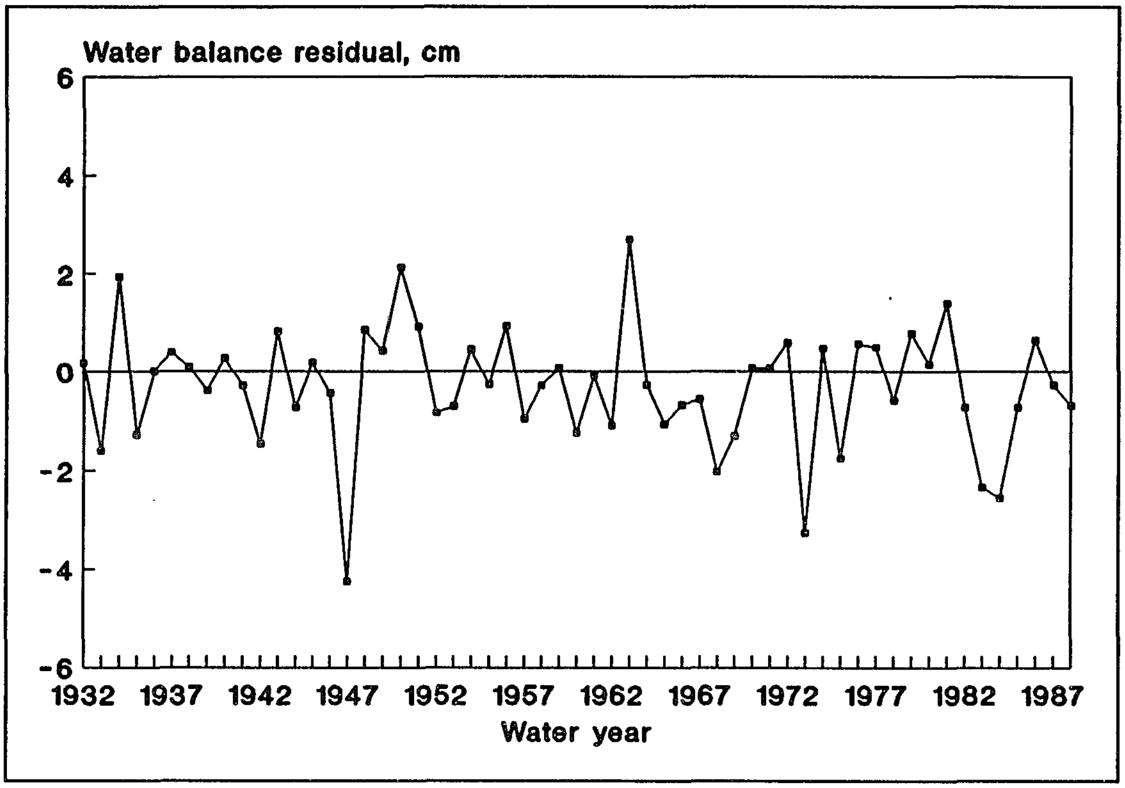

Figure 3.7. Time series graph of water balance residuals for the Mississippi basin at Tarbert Landing.

watershed. Another possibility to explain the discrepancy between discharge and modelled runoff could be that inaccurate discharge measurements were made at flood levels.

In 1954, actual discharge was the lowest over the years studied at 9.39 cm. The water balance calculates a modelled runoff that is slightly underestimated at 8.94 cm (an underestimation of 5%). In this case, additional discharge could have resulted from groundwater withdrawals or from a planned lowering of reservoir levels throughout the watershed.

In general, a pattern in the water balance residuals is not apparent for very wet or very dry years, yet some of the years with unusual climate conditions correspond to years with large absolute value residuals. Specific years with large water balance residuals are: 1947, 1968, 1973, 1983, 1984 (all negative); and 1934, 1950, 1956, 1963, 1981 (all positive). Three of these years were noted for remarkable flooding (1947, 1973, and 1983) and one for severe drought (1934).

A graphic indication of change in the water balance residuals over time can be illustrated by a cumulative residual mass curve in which the cumulative residuals are plotted against water year. If the magnitude of the water balance residuals remains relatively constant over a period of years, the data points follow a straight line. A change in the slope of the line is indicative of a change in the relationship between discharge and modelled runoff. Figure 3.8 clearly shows that the water balance residuals for this watershed have not been constant values over time. As illustrated in this figure, the cumulative residuals follow a fairly well-defined trend from 1932 through the mid-1960's. The year 1947 is an exception where the cumulative residuals shift dramatically; however, the cumulative residuals return to the established trend by 1950.

During the 1960's, the slope of the cumulative residuals decreases sharply over the remaining years, and the relationship between discharge and modelled runoff changes to produce larger and more variable residuals. Since the 1960's, the water balance estimates larger modelled runoff quantities than the discharge quantities actually measured. Possibly this change corresponds to the opening of the Old River Control Structure in July 1963 and the resulting gauge relocation from Red River Landing to Tarbert Landing. It should be noted that a change in the discharge/runoff relationship

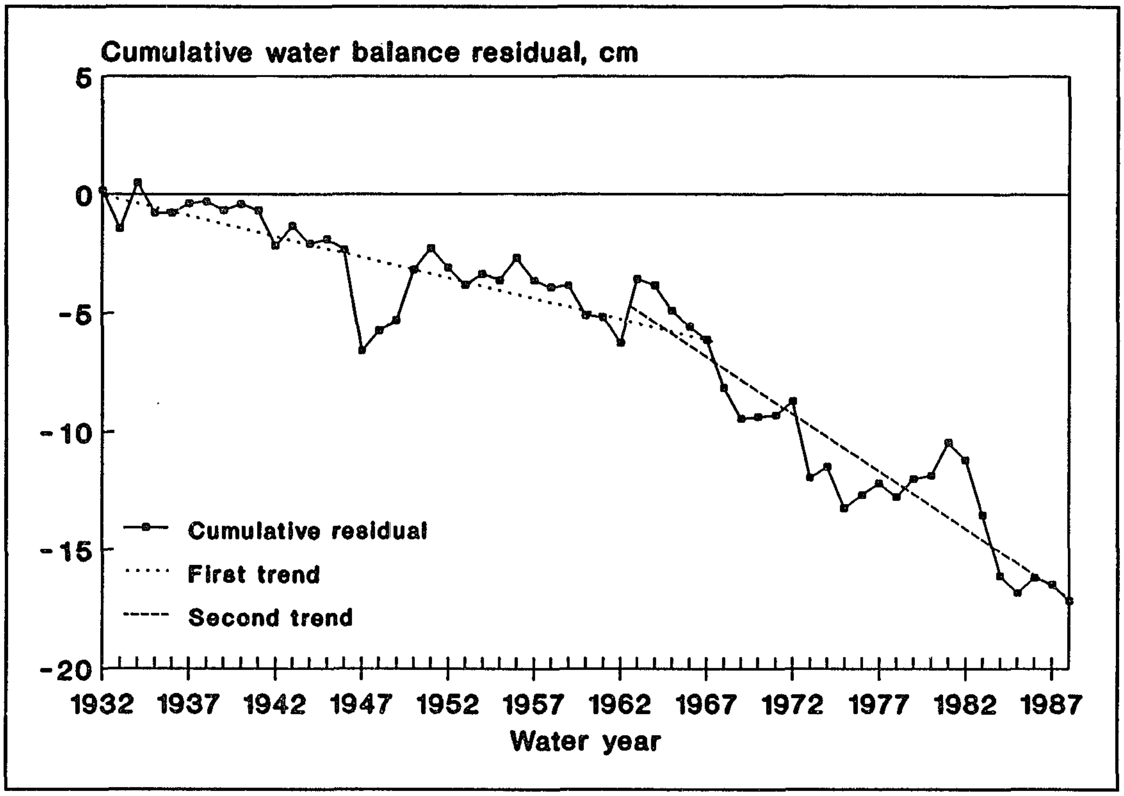

Figure 3.8. Time series graph of cumulative water balance residuals for the Mississippi basin at Tarbert Landing.

is obvious in the cumulative residuals graph; however, this change is not as obvious in the time series plot of the water balance residuals.

3.2 Linear Regression Results for the Mississippi Basin at Tarbert Landing

Least-squares linear regressions are used to determine which index (the water balance surplus or modelled runoff) is the best estimate of streamflow for the overall watershed. The F values and associated probabilities (Prob > F) for the regression models of surplus versus discharge and modelled runoff versus discharge are shown in Table 3.2. These results indicate that both regressions are significant at an α-level of 0.05. Further, large values of the adjusted coefficient of determination, r^2, reveal that both of the water balance indices explain greater than 95% of the total variability in annual discharge.

Table 3.2. Linear Regression Results of Water Balance Indices versus Discharge for the Mississippi Basin at Tarbert Landing

Index	F Value	Prob>F	Root MSE	Adj. r^2
Surplus*	1135.38	0.0001	1.1746	0.9530
Modelled runoff**	1117.76	0.0001	1.1721	0.9522

* $\text{Surplus}_t = -1.388 + 1.090\ \text{discharge}_t + \epsilon_t$
** $\text{Modelled runoff}_t = -1.184 + 1.079\ \text{discharge}_t + \epsilon_t$

There is very little difference between the root mean square errors (Root MSE) for these regressions; however, the adjusted r^2 is slightly larger for the regression based on surplus. This suggests that the water balance surplus (with no lag) may be a slightly better estimate of measured discharge than modelled runoff (with a one-month lag) for this drainage basin. Perhaps this is a result of a longer time-of-travel (two-month lag) within the Mississippi watershed. Using a two-month lag for the calculation of runoff could result in a better estimation of discharge and could therefore produce smaller water balance residuals. Or perhaps this is an artifact created in the data by summing discharge quantities from two gauging stations. For comparative purposes, the drainage basin at Vicksburg is also analyzed since records from only one gauge are used.

A plot of modelled runoff versus discharge is shown in Figure 3.9. Here, the least-squares regression line is only slightly different from a line with a slope of 1 (which would correspond to perfect correlation). This plot further demonstrates that

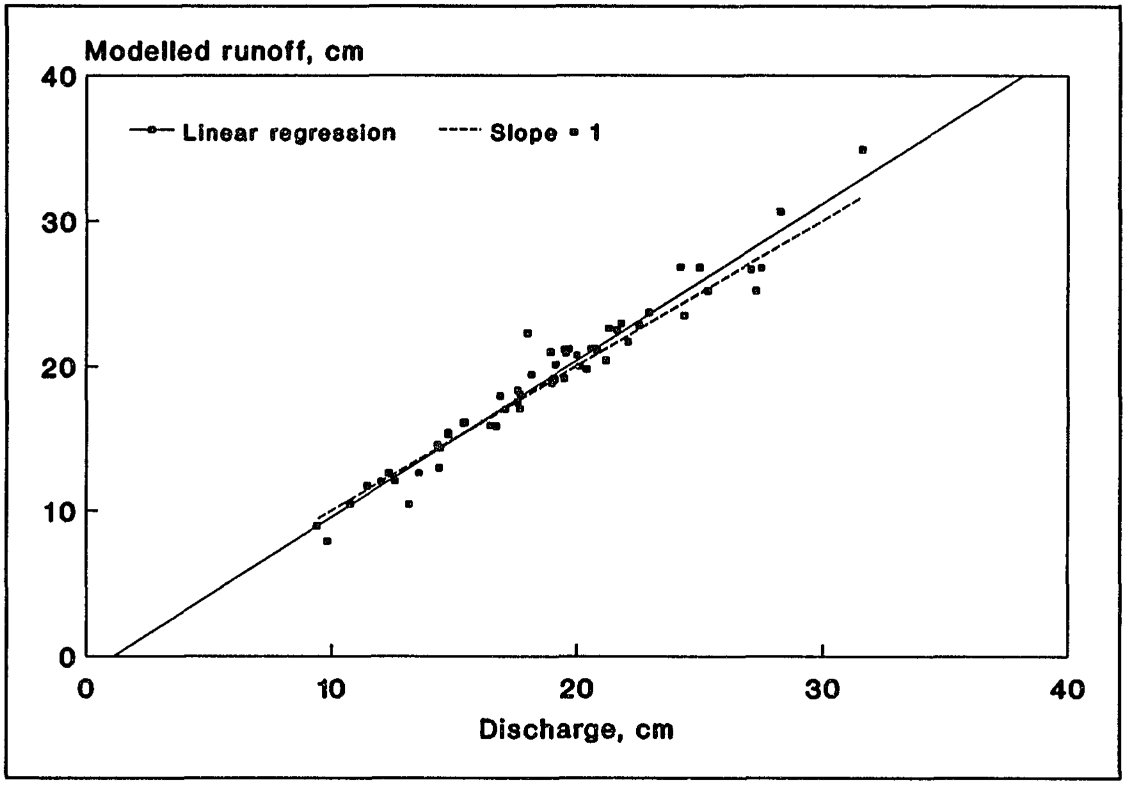

Figure 3.9. Modelled runoff versus annual discharge for the Mississippi basin at Tarbert Landing.

the water balance model fits extremely well for midrange values of annual discharge, but does not fit as well at lower and particularly at higher values of annual discharge.

As discussed earlier, certain assumptions pertaining to regression errors must be satisfied for linear regression analysis. The assumption of normality for the regression errors is tested with PROC UNIVARIATE. For both regressions, the results indicate that the regression errors are samples from normal distributions. The assumption of constant variance for the regression errors is tested by plotting the linear regression residuals against increasing values of discharge, against increasing values of the predicted value of the water balance index, and against time. These plots reveal that the assumption of constant variance is reasonable. Finally, the Durbin-Watson test statistics (D) calculated for surplus versus discharge (2.076) and for modelled runoff versus discharge (2.067) fall within the acceptance region, so the hypothesis of zero autocorrelation cannot be rejected. Thus, the basic assumptions are satisfied for both regressions, and the results are assumed to be valid.

3.3 Box-Jenkins Time Series Models for the Mississippi Basin at Tarbert Landing

An examination of the sample autocorrelation (SAC) function generated by the SAS procedure PROC ARIMA for the precipitation annual series reveals that the SAC has small autocorrelations at all lags (no spikes) and dies down fairly quickly. This type of behavior confirms that the time series is stationary over the years examined. The sample partial autocorrelation (SPAC) function also has small autocorrelations at all lags (no spikes) and dies down in a fashion dominated by damped sine wave oscillation. Thus, a tentative model for this series is one with no nonseasonal operator, i.e.,

$$y_t = \delta + a_t .$$

Parameter estimates and diagnostic information generated by SAS are used to finalize the model as follows:

1. The absolute t-value of 73.32 associated with the least squares point estimate of μ is considerably greater than 2, so the constant term δ should be included in the model. Since the model has no nonseasonal operator, then $\phi_p = 1$, and $\delta = \mu$. An estimation of the true mean for the series is 75.46 cm;
2. The overall standard error of 7.770 is the smallest error for all tentative models tested;

3. The final Box-Jenkins model adequately accounts for autocorrelation between time series observations as indicated by the large probability values for the Ljung-Box statistic. These are calculated at lag 6 (0.887), lag 12 (0.795), lag 18 (0.938), and lag 24 (0.877); and

4. Since there are no moving average or autoregressive operators in this model, there are no invertibility or stationarity conditions on the parameters.

Therefore, the Box-Jenkins model for annual precipitation as an average depth over the Mississippi basin above Tarbert Landing is

$$y_t = 75.46 + a_t,$$

where 75.46 cm is an estimation of the true mean of all possible realizations of the time series, and a_t is a random shock from a normal distribution. In effect, this is a white noise series which fluctuates around a mean of 75.46 cm.

The forecast for annual precipitation based on this model yields a 95% prediction interval of 60.23 cm to 90.69 cm. Thus, based on the 57-year period examined, a future observation will fall within this range with a 95% confidence. An examination of the historical annual precipitation series for comparative purposes shows that past values of precipitation fell outside of this 95% prediction interval for the years 1934 (58.20 cm), 1936 (58.55 cm), and 1973 (93.63 cm).

An examination of the behavior of the SAC and SPAC functions for the water balance surplus indicates that the SAC has a spike at lag 1 and cuts off after lag 1, and the SPAC has no spikes and dies down with damped sine wave oscillation. The behavior of the SAC indicates that the time series is stationary; therefore, data transformation is not required. A tentative nonseasonal moving average model of order 1 is selected, i.e.,

$$y_t = \delta + (1 - \theta_1 B)a_t,$$

where B is the backshift operator which shifts the subscript of a_t backwards in time by 1 lag. This model is called a moving average model because it uses the past random shock a_{t-1} in addition to the current random shock a_t to define the current time series value y_t. The calculation of t-values for μ and Θ_1 reveal that the t-value for Θ_1 is -1.94. As discussed earlier, an absolute t-value greater than 2 is required to produce a parsimonious model; however, 1.94 is nearly 2, and diagnostic checks reveal that including the parameter Θ_1 results in a model with a smaller overall standard error. The t-value for μ is 21.88, so the constant term should be included in the final model.

Thus, annual surplus in the overall watershed is adequately represented by the model

$$y_t = 19.10 + 0.2538a_{t-1} + a_t,$$

where 19.10 cm is an estimation of the true mean for the series, a_{t-1} is the random shock for the previous year, and a_t is the random shock for the current year. This model has an overall standard error of 5.277. In addition, the required invertibility conditions are satisfied since the absolute value of the Θ_1 parameter (-0.2538) is less than 1. The adequacy of this model is demonstrated by the high probability values obtained for the Ljung-Box statistic (0.760 at lag 6, 0.955 at lag 12, 0.997 at lag 18, and 0.900 at lag 24).

The 95% prediction interval forecast for the water balance surplus in the 58th year is 7.89 cm to 28.57 cm. By comparison, historical annual surplus quantities that fell outside of this interval occurred in 1934 (7.75 cm), 1973 (34.87 cm), and 1983 (30.64 cm). As shown earlier in the annual surplus maps, the surplus quantities that occurred during water years 1934 and 1973 were quite unusual as compared to mean annual surplus quantities.

Analysis of the SAC and SPAC for modelled runoff reveals that both functions have spikes at lag 1, and both functions cut off fairly abruptly. Thus, the behavior of the SAC indicates that the time series data are stationary. An examination of several tentative models reveals that a nonseasonal autoregressive model of order 1 provides the best fit for this series. The general equation for this type of model is

$$\left(1 - \phi_1 B\right) y_t = \delta + a_t,$$

where the backshift operator B acts on the time series observation y_t to shift the subscript backward in time by 1 lag. This model is called autoregressive because it defines the current time series value y_t as a function of the past time series value y_{t-1}. The final Box-Jenkins model for modelled runoff is

$$y_t = 14.02 + 0.2659 y_{t-1} + a_t,$$

where the standard error is 5.219. Further, the stationarity conditions for a first order autoregressive model are met since the absolute value of ϕ_1 (0.2659) is less than 1. The Ljung-Box probability values (0.669 at lag 6, 0.947 at lag 12, 0.998 at lag 18, and 0.909 at lag 24) indicate that nearly all of the autocorrelation between observations is accounted for in this model.

The 95% prediction interval forecast for modelled runoff in the 58th year is 8.08 cm to 28.53 cm. Once again, the historical years with a modelled runoff outside of this interval were 1934 (7.90 cm), 1973 (34.90 cm), and 1983 (30.63 cm).

The behavior of the SAC function for annual discharge indicates that the series is stationary over the years examined. The SAC has a spike at lag 1 and cuts off after lag 1, and the SPAC has a spike at lag 1 and dies down with damped sine wave oscillation. A nonseasonal moving average model of order 1 is selected for this series, and the final model is

$$y_t = 18.78 + 0.3621 a_{t-1} + a_t ,$$

which has a standard error of 4.554. The absolute value of Θ_1 (0.3621) satisfies the invertibility condition for a first order moving average model. Further, the Ljung-Box probability values indicate that this model is adequate (0.742 at lag 6, 0.855 at lag 12, 0.961 at lag 18, and 0.809 at lag 24).

A 95% prediction interval of 8.39 cm to 26.24 cm is forecast for annual discharge in the 58th year. Historically, the years that had actual discharge quantities outside of this interval were those with large flood occurrences: 1950 (27.31 cm); 1973 (31.64 cm); 1974 (27.10 cm); 1979 (27.52 cm); and 1983 (28.30 cm). Notably, in the years studied, there was not a single year with an annual discharge below the lower 95% limit. (The lowest annual discharge was 9.39 cm in 1954.) This is likely due to a fairly consistent base flow from groundwater supplies and also from reservoir releases during periods of low flow.

An examination of the behavior of the SAC calculated for the water balance residuals indicates that the time series is stationary. The SAC has a spike at lag 3, which indicates that the autocorrelation at lag 3 is statistically large, and cuts off after lag 3. The SPAC dies down in a fashion dominated by damped sine wave oscillation. It follows that a nonseasonal moving average model of the form

$$y_t = \mu + \left(1 - \theta_3 B^3\right) a_t$$

produces the best fit for the time series. Estimation of the model parameters results in the following final model,

$$y_t = 0.3034 - 0.2813 a_{t-3} + a_t ,$$

where the absolute value of Θ_1 (0.2813) satisfies the invertibility condition for a moving average model. This model has an overall standard error of 1.194. The Ljung-Box statistic probability values indicate that the model is adequate, although the probability values are not quite as large as those obtained for other models (0.264 at lag 6, 0.576 at lag 12, 0.800 at lag 18, and 0.938 at lag 24). However, an analysis of the residual sample autocorrelation (RSAC) function and the residual sample partial autocorrelation (RSPAC) function calculated for this model indicates that no spikes exist in the Box-Jenkins residuals.

A 95% prediction interval of -2.77 cm to 1.92 cm is forecast for the water balance residual in the 58th year. For comparative purposes, the historical years which produced water balance residuals outside of this interval were 1934 (1.93 cm), 1947 (-4.25 cm), 1950 (2.12 cm), 1963 (2.69 cm), and 1973 (-3.26 cm).

3.4 Summary of Findings for the Mississippi Basin at Tarbert Landing

Now, the statistical information given by these models can be interpreted in terms of the hydroclimatology of the drainage basin. According to the Box-Jenkins model developed for annual precipitation, the overall precipitation for the Mississippi basin can be described statistically as a randomly distributed, independent variable which fluctuates around a constant mean value. As indicated by the lack of autocorrelation between observations, precipitation in a specific year is not related to precipitation in any previous year. Therefore, no significant long-term trend exists in annual precipitation quantities for the Mississippi basin over the 57-year period.

As precipitation is converted to surplus, the Box-Jenkins model shifts from a purely random model to one that depends on the random shock generated in the previous year (since surplus values separated by a one-year lag are positively autocorrelated). Thus, the Box-Jenkins model developed for surplus emphasizes the hydrologic persistence usually apparent in surplus and streamflow. This means that a year of high surplus is more likely to be followed by a year of high surplus, while a year of low surplus is more likely to be followed by another year of low surplus. The model also reveals that the annual water balance surplus calculated for the basin contains no long-term trend over the years 1932 to 1988.

The annual modelled runoff also reveals a positive autocorrelation between values separated by a one-year lag. The time series for modelled runoff is stationary over the years studied.

Annual discharge for the Mississippi basin above Tarbert Landing reveals a positive autocorrelation between values separated by a one-year lag. It is interesting to note that the Box-Jenkins model for discharge more closely resembles the model for surplus (as opposed to modelled runoff) since both models utilize moving average operators. Thus, as suggested by the earlier linear regression results, the water balance surplus appears to be a better estimate of discharge for the overall Mississippi drainage basin. From 1932 to 1988, discharge is stationary with a constant mean, and there is no long-term trend toward increasing discharge for the watershed as a whole; however, recent years (1973, 1974, 1979, and 1983) have generated annual discharge quantities which have been unusually high. Notably, the flood of 1993 resulted in another year of high streamflow. If higher discharge quantities continue to occur over a number of years, a statistically significant trend could become apparent in this series over time.

The Box-Jenkins model developed for the water balance residuals reveals that residuals separated by a lag of three years are positively autocorrelated. (The subbasin analysis in the next chapter provides further insight to this three-year pattern.) This analysis shows that the annual water balance residuals (averaged over the watershed above Tarbert Landing) are stationary and have no significant long-term trend over the period studied. Therefore, any determination of whether land-use changes and river management have affected discharge for the overall basin is inconclusive for this basin.

3.5 Water Balance Results for the Mississippi Basin at Vicksburg

Table 3.3 compares the means, standard deviations, and test statistics for normality calculated for the Vicksburg-based annual series. The mean quantities calculated for the drainage basin at Vicksburg are slightly smaller than those presented for the Mississippi basin at Tarbert Landing. In terms of normality, the hypotheses that the series are samples from normal distributions cannot be rejected at an α-level of 0.05.

Table 3.3. Normality Test Results for the Mississippi Basin at Vicksburg Annual Series

Annual Series	Mean cm	Std. Dev. cm	W:Normal	Prob>W
Precipitation	73.58	7.42	0.9676	0.2590
Surplus	18.23	5.26	0.9862	0.9042
Modelled runoff	18.23	5.21	0.9866	0.9158
Discharge	18.08	4.57	0.9789	0.6489
Water balance residuals	-0.15	1.26	0.9723	0.3951

Annual precipitation and surplus for the Vicksburg data are graphed in Figure 3.10. Both of these series appear to be stationary with relatively constant variability. The mean annual precipitation for the Vicksburg data (at 73.58 cm) is approximately 2% less than the precipitation for the Tarbert Landing data. The range of annual precipitation quantities extends from 90.52 cm in 1973 to 56.69 cm in 1934.

Mean annual surplus for the Vicksburg drainage basin is 5% less than the mean surplus calculated for the Tarbert Landing basin. Here, surplus varies from a maximum of 33.39 cm in 1973 to a minimum of 6.95 cm in 1934.

Modelled runoff and discharge for Vicksburg are compared in Figure 3.11. On the whole, modelled runoff agrees with actual discharge fairly well, except for a period from 1982 to 1986 where actual discharge was lower than that estimated by the model.

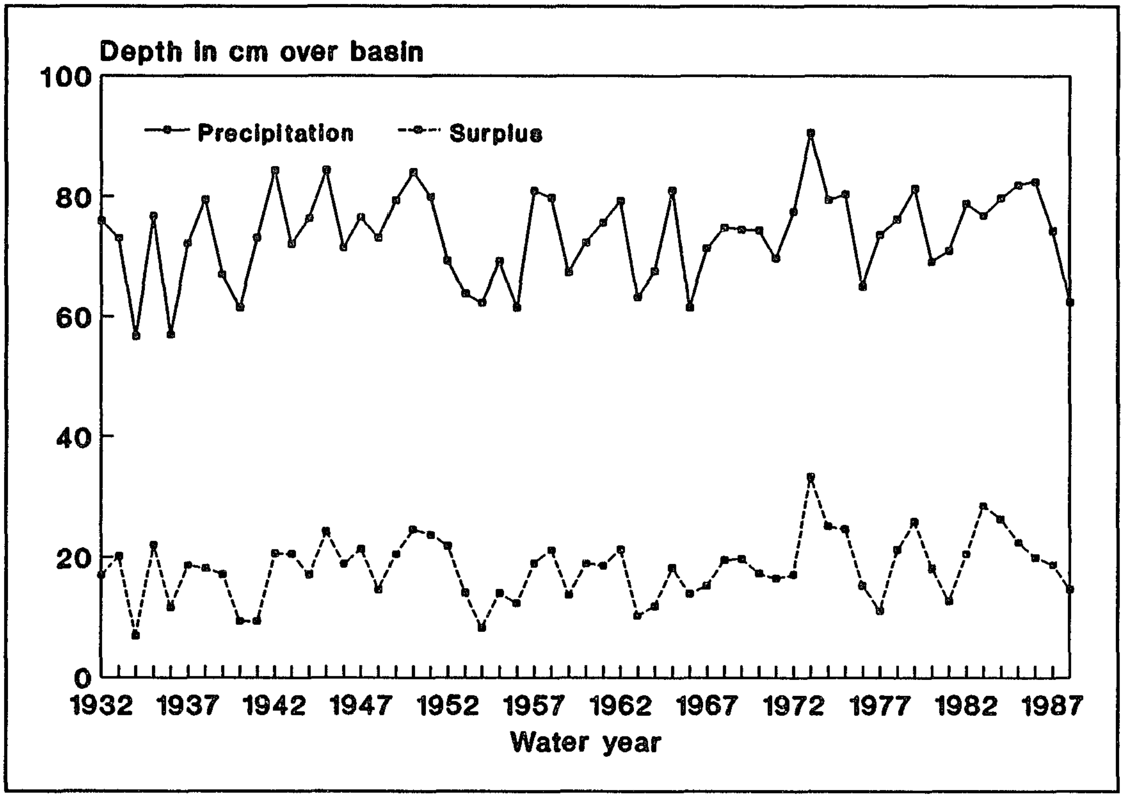

Figure 3.10. Times series graph of annual precipitation and water balance surplus for the Mississippi basin at Vicksburg.

The mean annual discharge for the Vicksburg-based data (at 18.08 cm) is 4% smaller than the Tarbert Landing discharge since discharge from the Red River basin is not included. Further, the range of discharge values is smaller for the Vicksburg data (21.37 cm as opposed to 22.25 cm for the Tarbert Landing data), and the standard deviation of 4.57 cm shows that the discharge measured at Vicksburg is slightly less variable than the Tarbert Landing discharge (4.85 cm). It is a reasonable assumption that the summing of data from two gauges to create the Tarbert Landing-based data set introduces greater variability into the discharge data. Therefore, the Vicksburg-based data should provide a more consistent database for statistical analysis.

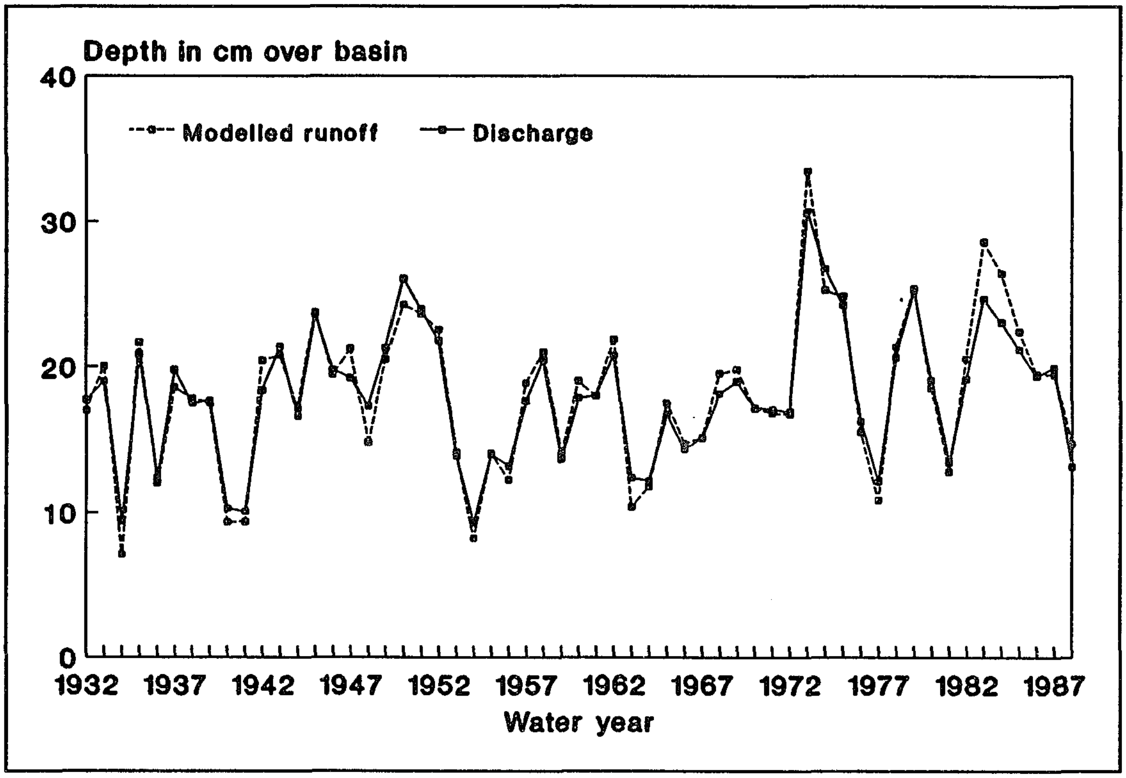

Figure 3.11. Time series graph of annual discharge and modelled runoff for the Mississippi basin at Vicksburg.

Figure 3.12 is a graph of the water balance residuals calculated for the Vicksburg data. The average magnitude of the water balance residuals for this basin (based on absolute values of the residuals) is 0.95 cm (approximately 5% of the mean discharge at Vicksburg). The residuals fluctuate around a mean value of -0.15 cm with a standard deviation of 1.26 cm. The maximum water balance residual occurs in 1948 (2.48 cm), and the minimum water balance residual occurs in 1983 (-3.96 cm).

In general, the graphs for the Vicksburg data vary little from those presented for the Tarbert Landing data except for the reduction in discharge variability. However, the cumulative residual mass curve for the Vicksburg-based data is substantially different. The Vicksburg-based cumulative water balance residuals (shown in Figure

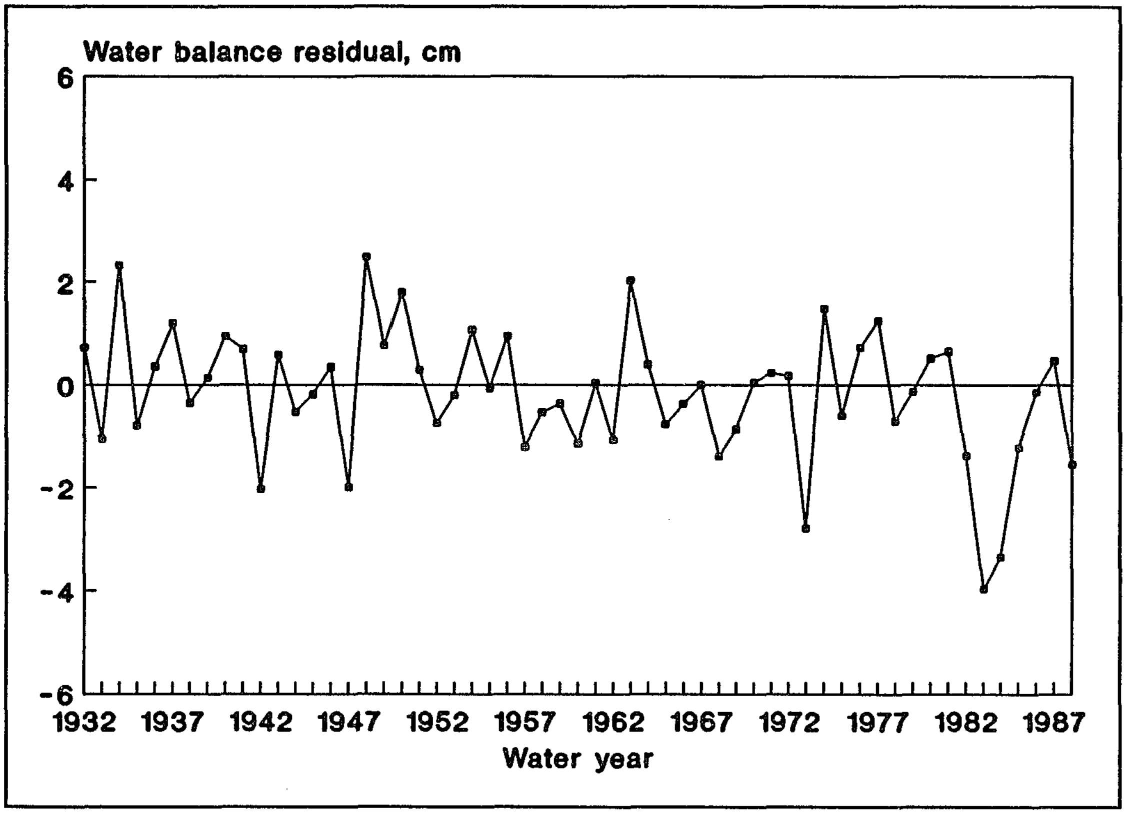

Figure 3.12. Time series graph of water balance residuals for the Mississippi basin at Vicksburg.

3.13) follow a general trend of underestimating discharge from 1932 until the early 1950's. During the early 1950's, the cumulative residuals shift so that the water balance overestimates discharge until the late 1970's or so. Then, in 1982, a dramatic shift occurs so that modelled runoff is consistently much greater than actual discharge. This shift is also apparent in the time series graph of the water balance residuals (as a large negative peak) and in the graph of annual discharge and modelled runoff (as a large discrepancy in values between 1982 to 1985). Quite possibly these trends correspond to gauge relocations or revised gauge-discharge curves as postulated earlier for the Tarbert Landing data. If so, this could cast some doubt on the accuracy and precision of discharge data over extended periods of time. As mentioned earlier, any inaccuracies in measured discharge appear as water budget residuals in this study. Thus, any effects of land-use change and river management on discharge may be masked by other factors affecting the water balance residuals.

3.6 Linear Regression Results for the Mississippi Basin at Vicksburg

The results from linear regressions comparing the water balance surplus and modelled runoff with discharge are shown in Table 3.4. The values of the adjusted r^2 are slightly lower for these regressions as compared to those obtained for the Tarbert Landing regressions, which means that the water balance indices explain slightly less of the total variability in annual discharge (approximately 95%) for this basin. Yet, the water balance model quite effectively estimates discharge as shown in the graphs of modelled runoff versus discharge. Since the values of the adjusted r^2 are higher and the root MSE values are lower for the regression based on modelled runoff, it is apparent that modelled runoff estimates discharge better than the water balance surplus in this watershed. Therefore, it appears that the water balance model fits this drainage basin better than the Tarbert Landing basin.

Table 3.4. Linear Regression Results of Water Balance Indices versus Discharge for the Mississippi Basin at Vicksburg

Index	F Value	Prob > F	Root MSE	Adj. r^2
Surplus*	992.173	0.0001	1.2159	0.9465
Modelled runoff**	1058.676	0.0001	1.1687	0.9497

* $\text{Surplus}_t = -2.000 + 1.119\ \text{discharge}_t + \epsilon_t$

** $\text{Modelled runoff}_t = -1.855 + 1.111\ \text{discharge}_t + \epsilon_t$

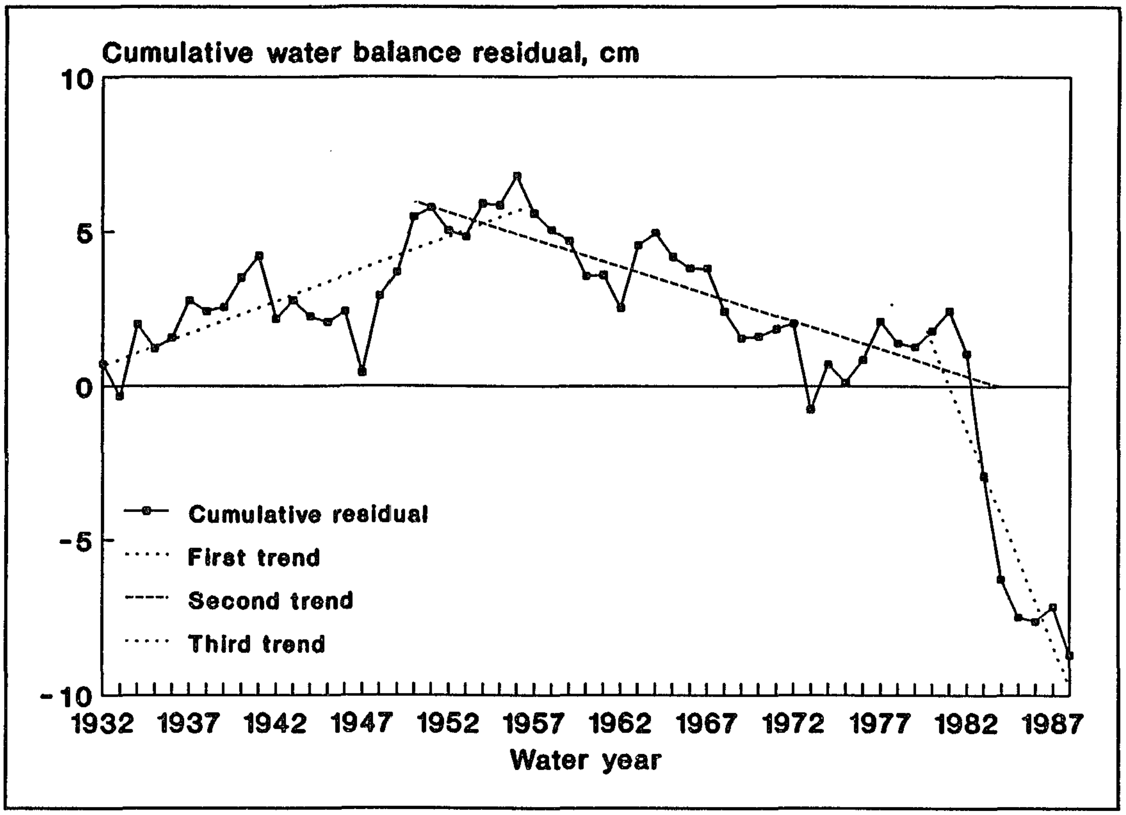

Figure 3.13. Time series graph of cumulative water balance residuals for the Mississippi basin at Vicksburg.

Figure 3.14 contains a plot of modelled runoff versus discharge for the Vicksburg-based data. As seen earlier in the Tarbert Landing-based data, the water balance model does not estimate discharge as accurately during extreme conditions and particularly at higher discharges.

To determine whether the regression results are valid, the assumptions for linear regression analysis are examined as follows. Results from the SAS procedure PROC UNIVARIATE for the regression error terms indicate that the errors are normally distributed in both regressions. Plots of the linear regression errors reveal that the assumption of constant variance is reasonable. Further, an analysis of first order

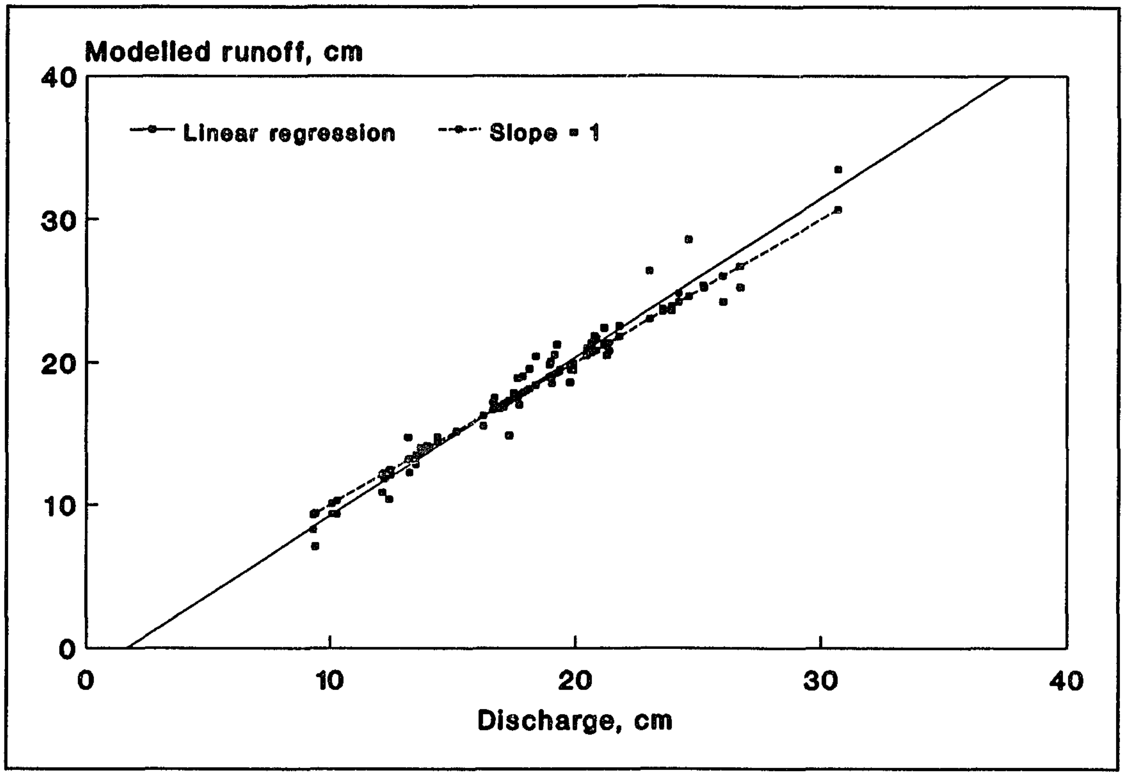

Figure 3.14. Modelled runoff versus annual discharge for the Mississippi basin at Vicksburg.

autocorrelation indicates that the Durbin-Watson test statistics fall within the acceptance region (1.819 for surplus versus discharge and 1.704 for modelled runoff versus discharge). As such, the assumption of independent regression errors cannot be rejected, and overall, the assumptions for linear regression are satisfied.

3.7 Box-Jenkins Time Series Models for the Mississippi Basin at Vicksburg

The SAC for the Vicksburg-based annual precipitation has no spikes and dies down fairly quickly, which indicates that the time series is stationary. The SPAC function has no spikes and also dies down fairly quickly. Diagnostic checks verify that the final Box-Jenkins model for annual precipitation in this watershed is one with no nonseasonal operator, specifically,

$$y_t = 73.58 + a_t.$$

The standard error for this model is 7.423, and there are no invertibility or stationarity conditions for this model. The Ljung-Box probability values are quite high (0.935 at lag 6, 0.819 at lag 12, 0.940 at lag 18, and 0.919 at lag 24) which indicate that the model is adequate.

The 95% prediction interval forecast for precipitation is 59.03 cm to 88.13 cm. Historically, years with precipitation outside of this interval were 1934 (56.69 cm), 1936 (56.97 cm), and 1973 (90.52 cm).

Both the SAC and SPAC for the Vicksburg-based water balance surplus have spikes at lag 1, and both functions die down fairly quickly. Thus, the time series is stationary over the years studied. Evaluation of several tentative models indicates that annual surplus is adequately represented by a nonseasonal autoregressive model of order 1. The final Box-Jenkins model is

$$y_t = 13.31 + 0.2688 y_{t-1} + a_t,$$

where the standard error is 5.112. The stationarity conditions are satisfied since ϕ_1 (0.2688) is less than 1. In addition, the Ljung-Box statistic has large probability values which indicate that nearly all of the autocorrelation between observations is accounted for by this model (0.721 at lag 6, 0.940 at lag 12, 0.996 at lag 18, and 0.885 at lag 24).

The 95% prediction interval forecast for surplus in the 58th year is 7.22 cm to 27.25 cm. In comparison, surplus quantities in 1934 (6.95 cm), 1973 (33.39 cm), and 1983 (28.56 cm) fell outside of this interval.

The SAC and the SPAC calculated for modelled runoff have spikes at lag 1 and cut off. The time series is stationary, and parameter estimates indicate that this series is best represented by an autoregressive model of order 1. The appropriate Box-Jenkins model for this series,

$$y_t = 13.18 + 0.2759y_{t-1} + a_t,$$

has an overall standard error of 5.057. The Ljung-Box statistic probability values are 0.707 at lag 6, 0.956 at lag 12, 0.998 at lag 18, and 0.914 at lag 24. Further, the stationarity conditions are satisfied since ϕ_1 (0.2759) is less than 1.

A 95% prediction interval of 7.33 cm to 27.15 cm is forecast for modelled runoff in the 58th year. Historically, the modelled runoff calculated for the years 1934, 1973, and 1983 fell outside of these limits at 7.08 cm, 33.46 cm, and 28.57 cm, respectively.

The annual discharge for Vicksburg is a stationary time series as indicated by the behavior of the SAC function (which has a spike at lag 1 and cuts off). The SPAC also has a spike at lag 1. Estimation of model parameters and diagnostic checking reveals that this series is adequately modelled by a nonseasonal autoregressive model of order 1, i.e.,

$$y_t = 11.67 + 0.3528y_{t-1} + a_t,$$

with an overall standard error of 4.325. The stationarity condition for a first-order autocorrelation model is satisfied since the estimate of ϕ_1 (0.3528) is less than 1. The probability values for the Ljung-Box statistic are 0.612 at lag 6, 0.858 at lag 12, 0.957

at lag 18, and 0.776 at lag 24; therefore, the model adequately accounts for autocorrelation between observations.

A 95% prediction interval of 7.84 cm to 24.79 cm is forecast for annual discharge in the 58th year. Historically, the annual discharge quantities for 1950 (26.02 cm), 1973 (30.67 cm), 1974 (26.71 cm), and 1979 (25.22 cm) exceeded the upper bound of this interval. Again, there were no years with an annual discharge smaller than the lower bound of the prediction interval.

The SAC for the Vicksburg-based water balance residuals indicates that the series is stationary, since both the SAC and SPAC die down fairly quickly with no spikes. The t-value calculated for μ is 0.91, which indicates that μ should not be included in a parsimonious model. Accordingly, the final model is simply

$$y_t = a_t .$$

As a result, the water balance residuals for this basin are a purely random, white noise series drawn from a normal distribution. The overall standard error for this model is 1.263. High probability values for the Ljung-Box statistic support the adequacy of this model (0.740 at lag 6, 0.909 at lag 12, 0.890 at lag 18, and 0.943 at lag 24).

The 95% prediction interval forecast for future values of the water balance residual is -2.48 cm to 2.48 cm. Specific years with historical water balance residuals at or outside of these limits were 1948 (2.48 cm), 1973 (-2.79 cm), 1983 (-3.96 cm), and 1984 (-3.35 cm).

3.8 Summary of Findings for the Mississippi Basin at Vicksburg

In the Mississippi basin above Vicksburg, as in the Mississippi basin above Tarbert Landing, the Box-Jenkins model indicates that precipitation is a white noise series. In other words, precipitation in one year is not related to precipitation in another year. The Box-Jenkins model for the Vicksburg-based data has a smaller standard error and a smaller 95% prediction interval forecast.

The models for annual surplus, modelled runoff, and discharge also reflect a one-year positive autocorrelation which is characteristic of hydrological data. Again for these models, the standard errors and 95% prediction intervals are smaller for the Vicksburg-based data than for the Tarbert Landing-based data. Further, it is interesting to note that the Box-Jenkins models derived for surplus, modelled runoff, and discharge are quite similar since they are based on autoregressive operators. This provides additional evidence that the water balance produces extremely good estimates of annual streamflow for the Vicksburg watershed.

The Box-Jenkins model for the Vicksburg-based water balance residuals is quite striking in its implications. This model indicates that the water balance estimates annual discharge for this basin so accurately that a random error series with a mean of zero is the only residual. Thus, over the past 57 years, any effects on discharge due to human modification are either too small to detect on such broad scales (both temporal and spatial), or competing, such that the net result appears as though there has been no effect on discharge. The years 1948, 1973, 1983, and 1984 have water balance residuals that are statistically significant, so perhaps some specific explanations can be identified for these years.

3.9 Chapter Summary

This chapter has presented statistical data, time series graphs, linear regression comparisons, and Box-Jenkins models for annual time series in the overall Mississippi drainage basin. The water balance indices of surplus and modelled runoff estimate annual discharge with extremely successful results since the model is able to account for approximately 95% of the annual variance in streamflow.

In regard to the Box-Jenkins models developed for each time series, annual precipitation is purely random over the years studied. Annual surplus, modelled runoff, and discharge quantities are strongly autocorrelated based on a one-year lag. The water balance residuals are purely random for the Mississippi basin at Vicksburg, while the residuals for the Mississippi basin at Tarbert Landing demonstrate a three-year pattern of positive autocorrelation.

All of these series are stationary with no indication of long-term trends over the time period 1932 to 1988. Therefore, any changes in annual precipitation or surplus due to climatic change or any effects on annual discharge due to regulation and/or land-use changes have been either insignificant or masked by other factors. Since it is possible that these changes and effects cannot be detected over such a large geographic area, two subbasins within the Mississippi drainage basin are analyzed in the next chapter.

CHAPTER 4

THE MISSOURI AND OHIO-TENNESSEE BASINS

4.0 Chapter Objectives

A major objective of this chapter is to evaluate how accurately the water balance model estimates annual streamflow for two subbasins of the Mississippi drainage basin: the Missouri and the Ohio-Tennessee. As in the previous chapter, the water balance indices of surplus and modelled runoff are compared with actual discharge data to evaluate the effectiveness of the water balance model over a smaller geographic area.

In order to determine whether the hydroclimatology of the watersheds has changed over time, Box-Jenkins time series models are derived for nonseasonal (water year) time series data. In addition, the portion of the annual Mississippi discharge supplied by each subbasin is evaluated to determine whether changes have occurred during the 57-year period of study. Again, the statistical and physical implications of each model are discussed.

Another objective of this chapter is to determine if human modifications have affected discharge from the Missouri and Ohio-Tennessee subbasins. Box-Jenkins time series analysis is used to evaluate the residuals from the water balance model to determine whether changes have occurred.

4.1 Water Balance Results for the Missouri Basin at Hermann

The characteristics of the annual series for the Missouri basin are quite different from those calculated for the overall Mississippi basin since the water balance model almost consistently underestimates the amount of runoff generated within the drainage area. Table 4.1 contains the means, standard deviations, and normality test results (W:Normal and Prob>W) for the annual series. In this watershed, the calculated surplus, modelled runoff, and the water balance residuals are not samples from a normal population. Because surplus and modelled runoff cannot be less than zero by definition, the skewness values for surplus (0.8405) and modelled runoff (0.7924) indicate that these distributions are skewed to the right. Further, since the distributions contain no negative values, the kurtosis values for surplus (-0.0608) and modelled runoff (-0.2351) reveal that the tails are lighter than those of a normal distribution. The distribution for the water balance residuals is skewed to the left (-0.9739). Since the residuals are relatively large, the kurtosis for the water balance residuals (0.9485) indicates that the tails of this series are heavier than those of a normal distribution.

Table 4.1. Normality Test Results for the Missouri Basin at Hermann Annual Series

Annual Series	Mean cm	Std. Dev. cm	W:Normal	Prob>W
Precipitation	48.56	6.35	0.9735	0.4365
Surplus	3.50	2.45	0.9077	0.0002
Modelled runoff	3.50	2.41	0.9135	0.0004
Discharge	5.14	1.83	0.9637	0.1757
Water balance residuals	1.65	0.96	0.9318	0.0040

Figure 4.1 contains a graph of annual precipitation and calculated surplus for the Missouri basin. (Note that the y-axis scales used for the Missouri subbasin figures are the same as those used for the overall Mississippi basin.) Mean annual precipitation over the watershed is 48.56 cm (with a standard deviation of 6.35 cm). The range

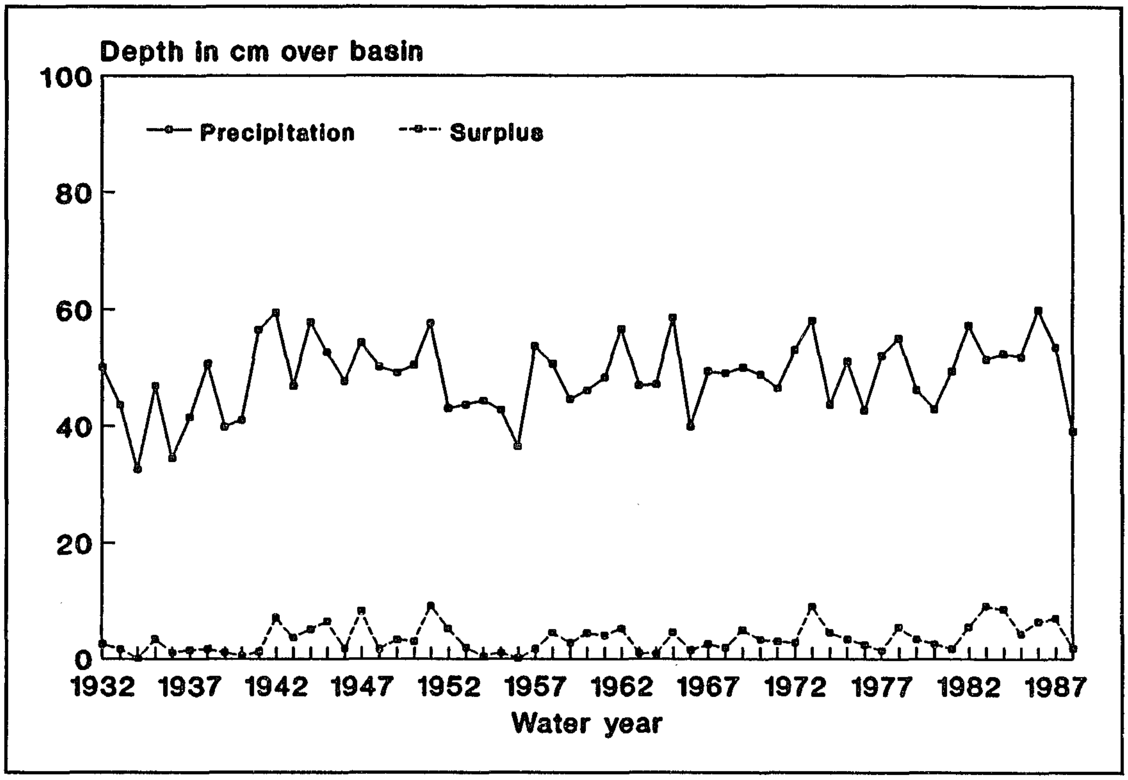

Figure 4.1. Time series graph of annual precipitation and water balance surplus for the Missouri basin at Hermann.

extends from a maximum of 59.89 cm in 1986 (123% of the mean precipitation) to a minimum of 32.52 in 1934 (67% of the mean precipitation). Variability in precipitation is relatively larger for this subbasin (the standard deviation is 13% of the mean precipitation) as compared to the Mississippi basin at Vicksburg (where the standard deviation is only 10% of the mean precipitation). The semiarid climate of the Missouri basin is evident in the relatively small quantity of annual surplus available. As noted earlier, several climate divisions in this drainage basin typically have annual surplus values of zero. Mean annual surplus is only 3.50 cm (7% of the mean precipitation) averaged over the drainage basin area. The maximum annual surplus is 9.21 cm in 1951 (19% of the mean precipitation), while the minimum annual surplus is 0.06 cm in 1934 (0.1% of the mean precipitation). Thus, variability of surplus in this watershed is more extreme than in the overall Mississippi basin. As a result, floods are more damaging and droughts are more devastating when they occur.

Figure 4.2 is a graph of discharge and modelled runoff for the Missouri basin. The water balance model underestimates annual streamflow in this basin by an average of 32% over the years studied. In 1973, the modelled runoff of 8.79 cm only slightly underestimates (by 5%) the actual discharge value of 9.24 cm. Comparatively, in 1934 the modelled runoff of 0.09 cm drastically underestimates (by 95%) the actual discharge value of 1.96 cm.

The mean annual discharge averaged over the area of the drainage basin is 5.14 cm (with a standard deviation of 1.83 cm). Annual discharge ranges from a maximum of 9.24 cm in 1973 (180% of the mean discharge) to a minimum of 1.96 cm in 1934 (38% of the mean discharge). Since the smallest annual discharge values occurred during the dust bowl years of the 1930's and the little dust bowl years of the 1950's, the time series for annual discharge visually appears to have a gradual upward trend over the 57-year period. This appearance of increasing discharge within the basin is further emphasized in more recent decades by high variability and frequent major flood occurrences. The extension of this time series plot to include 1993 would further exaggerate the appearance of increasing discharge over time.

To determine whether the flow dynamics within the overall drainage basin have changed over time, the discharge contributed by the Missouri basin is calculated as a fraction of the total Mississippi River discharge measured at Vicksburg. Figure 4.3 is

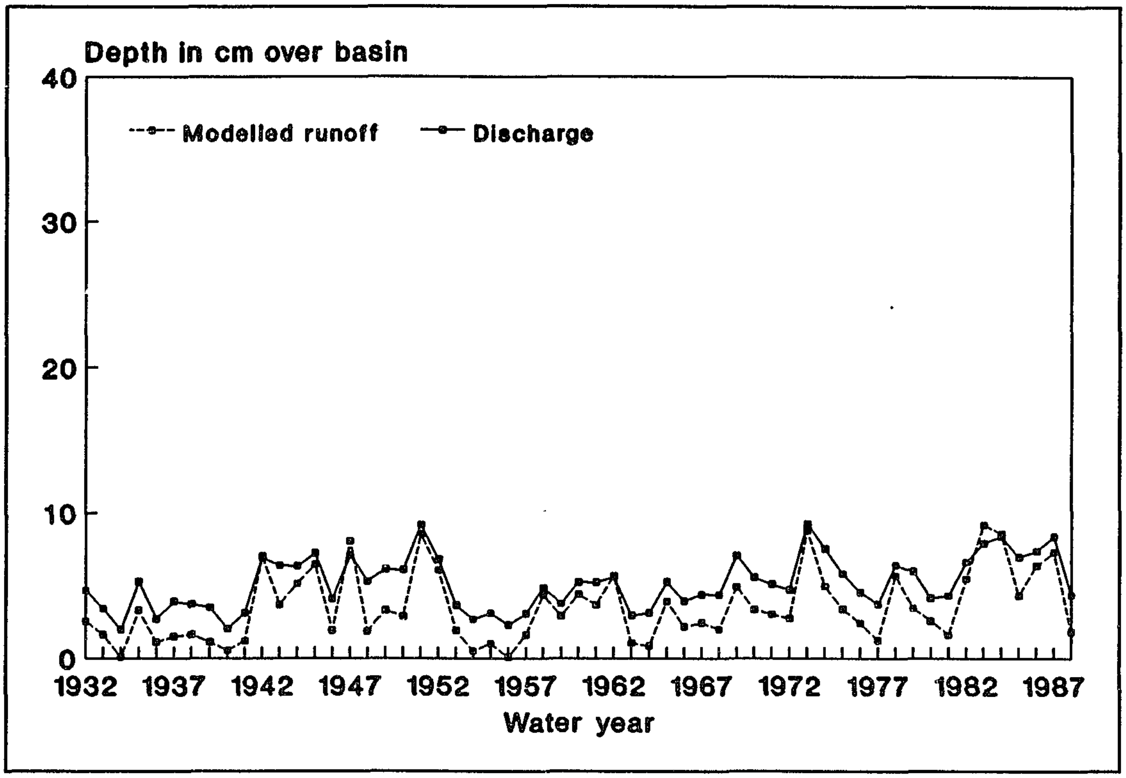

Figure 4.2. Time series graph of annual discharge and modelled runoff for the Missouri basin at Hermann.

a graph which illustrates the Missouri discharge fraction over the years 1932 to 1988. Even with the largest drainage area of any subbasin, the Missouri subbasin produces only 13.20% of the total Mississippi basin streamflow on average. The Missouri discharge fraction ranges from a maximum in 1987 (19.75%) to a minimum in 1956 (8.18%). The time series shows a period of smaller discharge fractions from 1953 to 1967 which coincides with the construction of a series of flood control dams within the basin (Keown, et al. 1986). It is reasonable to assume that the Missouri subbasin contribution to total flow would be diminished as large reservoirs were filled upstream.

The average magnitude of the water balance residuals in the Missouri subbasin (based on the absolute value of the residuals) is 1.74 cm (nearly 34% of the mean

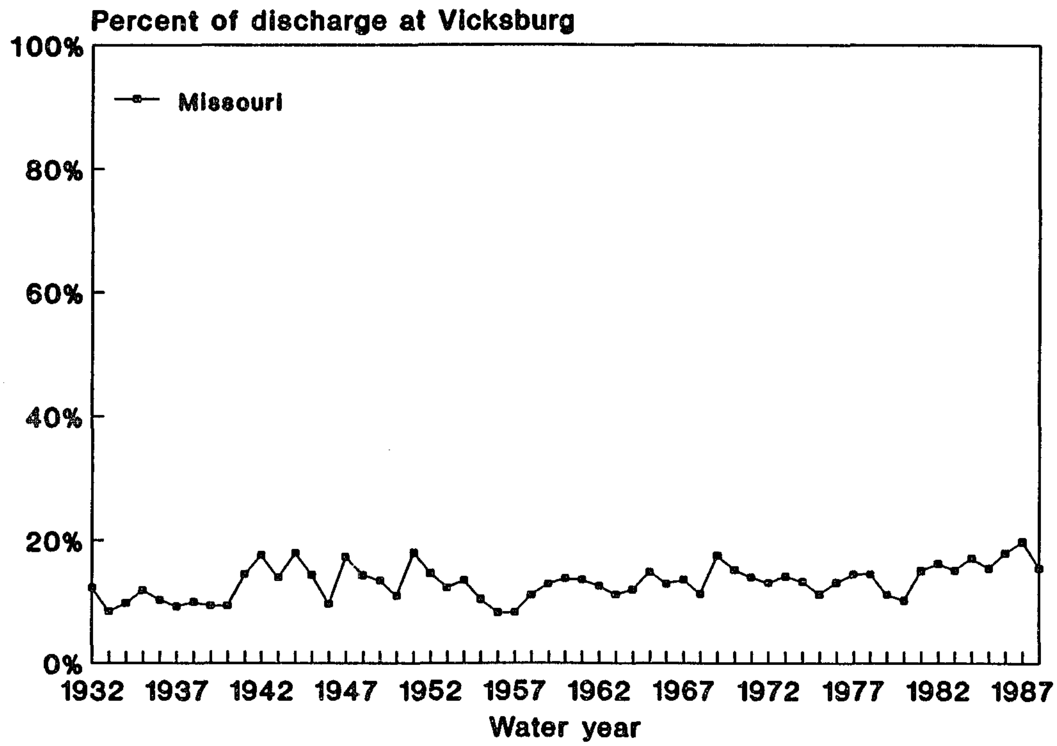

Figure 4.3. Time series graph of the discharge fraction for the Missouri basin at Hermann.

annual discharge). Because measured discharge is nearly always larger than the runoff estimated by the water balance, the water balance residuals are primarily positive values that fluctuate around a mean of 1.65 cm with a standard deviation of 0.96 cm. Comparatively, the range of residual values is smaller for this watershed than for the overall Mississippi basin and extends from 3.41 cm in 1948 to -1.24 cm in 1983.

As shown in Figure 4.4, only five years have negative water balance residual values: 1942 (-0.10 cm); 1947 (-0.97 cm); 1962 (-0.09 cm); 1983 (-1.24 cm); and 1984 (-0.19 cm). Since the years 1983 and 1984 also correspond to years of high discharge, it appears that for this subbasin the water balance model is more accurate

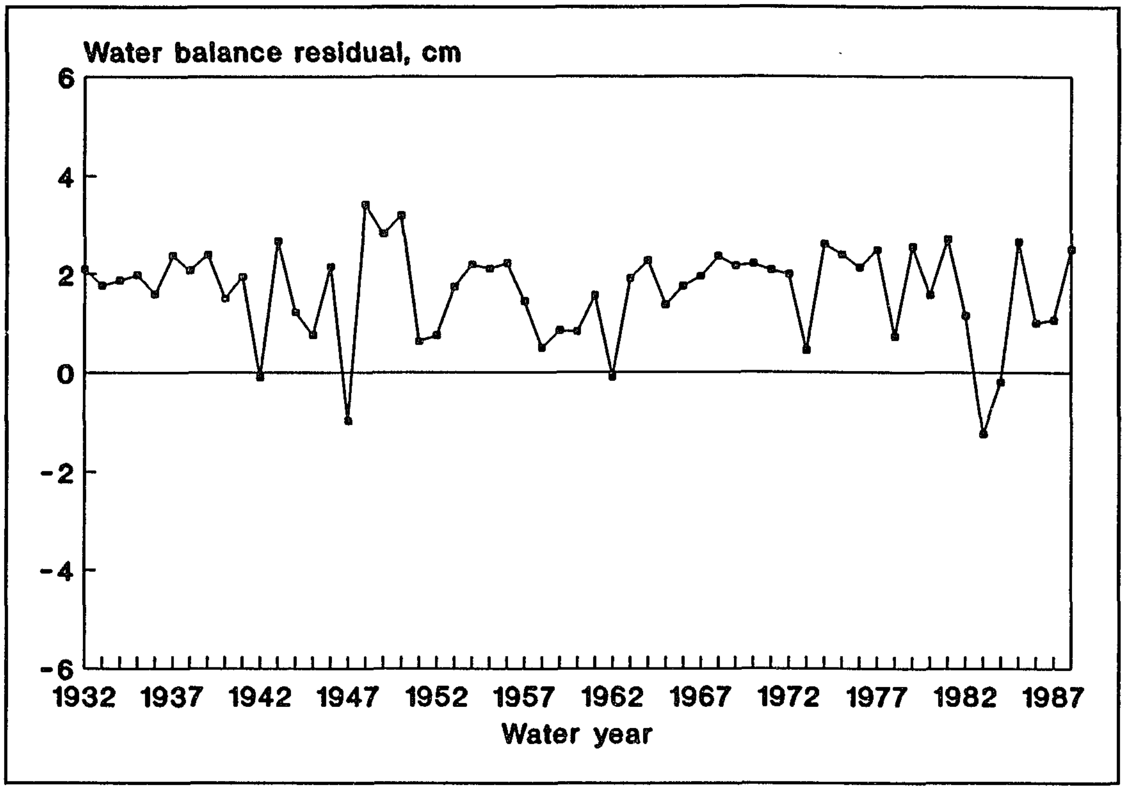

Figure 4.4. Time series graph of water balance residuals for the Missouri basin at Hermann.

for years with high discharge than for years with low discharge. An interesting feature of this graph is the relatively flat water balance residual curve from 1964 to 1972, when variability is considerably damped.

The most likely reason for the inaccuracies in modelled runoff is that the divisional climate data do not account for orographic precipitation that falls mostly as snow in the Rocky Mountains. Bradley (1976) demonstrated that in mountainous areas streamflow data are a better indication of the quantity of rainfall than precipitation data since precipitation gauges generally are biased to lower elevations. Also, Stockton and Jacoby (1976) determined that the main component of streamflow in the Colorado River basin was snowmelt runoff from mountainous areas. According to their estimation, 85% of the Colorado River discharge originated from only 15% of the basin's area. Changnon, et al. (1990) concluded that in 11 of 14 Rocky Mountain drainage basins, accumulated snowpack had a better relationship to streamflow than winter precipitation. As discussed earlier, Dolph and Marks (1992) showed that the incorporation of snow data into the precipitation record increased precipitation by approximately 40% in the western U.S. In addition, some of the discrepancy between measured discharge and modelled runoff may be due to groundwater withdrawals and/or reservoir storage within the watershed since neither of these factors are taken into account by the water balance model.

The cumulative residual mass curve for the Missouri basin indicates that the magnitude of the water balance residuals is nearly constant over time as shown in Figure 4.5. This graph suggests several possibilities:

1. The relationship between discharge and modelled runoff has not changed over the years studied; or
2. Any effect on measured discharge due to human modification is so small by comparison that it cannot be detected; or
3. Contemporaneous effects on discharge may be competing such that the net result appears to be no effect at all.

Further, it appears that gauge relocations and revisions of the stage-discharge curve at Hermann have not resulted in a significant effect on the cumulative water balance residuals over the 57-year period.

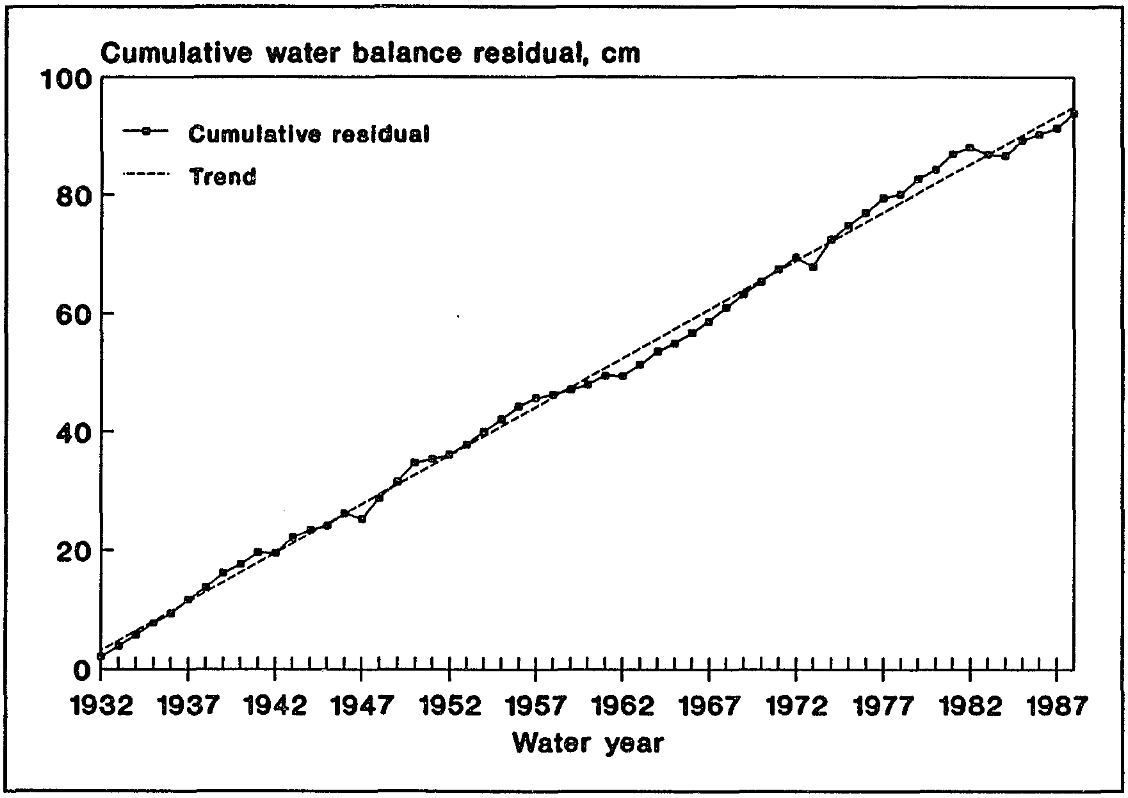

Figure 4.5. Time series graph of cumulative water balance residuals for the Missouri basin at Hermann.

4.2 Linear Regression Results for the Missouri Basin at Hermann

Results of linear regression analyses for this subbasin are shown in Table 4.2. The adjusted r^2 values indicate that greater than 86% of the variability in annual discharge is explained by the water balance indices. In the Missouri subbasin, the modelled runoff is a better index than surplus since it explains slightly more of the variation in discharge and has a considerably smaller root MSE. The one-month lag used to calculated modelled runoff appears to be a good approximation of the actual lag time for the overall basin. Thus, the water balance is reasonably effective here despite the general underestimation of streamflow by the model.

Figure 4.6 is a plot of modelled runoff versus discharge. As suggested earlier, it appears that the water balance is more effective at estimating larger discharge quantities than estimating smaller discharge quantities.

The assumptions for linear regression are verified as follows. The linear regression errors are determined to be samples from a normal distribution by analyzing results from the SAS procedure PROC UNIVARIATE. Plots of the linear regression errors against various criteria indicate that the assumption of constant variance for the regression errors is reasonable. Finally, the Durbin-Watson test statistics for surplus versus discharge (1.903) and modelled runoff versus discharge (1.797) fall within the acceptance region for zero autocorrelation. Therefore, the linear regression results obtained for this watershed are assumed to be valid.

Table 4.2. Linear Regression Results of Water Balance Indices versus Discharge for the Missouri Basin at Hermann

Index	F Value	Prob>F	Root MSE	Adj. r^2
Surplus*	347.542	0.0001	0.9134	0.8609
Modelled runoff**	366.123	0.0001	0.8771	0.8670

* $\text{Surplus}_t = -2.913 + \text{discharge}_t + \epsilon_t$

** $\text{Modelled runoff}_t = -2.882 + 1.229\ \text{discharge}_t + \epsilon_t$

4.3 Box-Jenkins Time Series Models for the Missouri Basin at Hermann

An examination of the sample autocorrelation (SAC) function for annual precipitation in the Missouri drainage basin reveals that the SAC has a spike at lag 3 and cuts off after lag 3; therefore, annual precipitation for the Missouri subbasin is stationary over the time period of this study. The sample partial autocorrelation

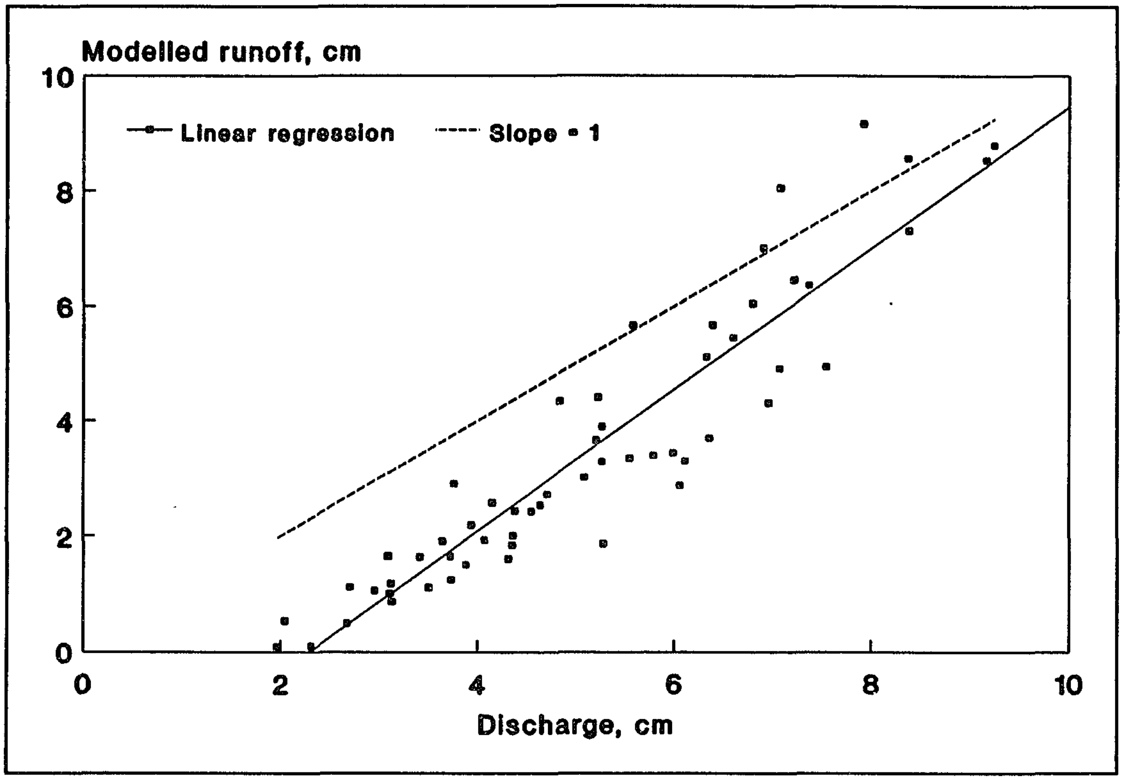

Figure 4.6. Modelled runoff versus annual discharge for the Missouri basin at Hermann.

(SPAC) function contains no spikes and dies down in a fashion dominated by exponential decay. Calculation of model parameters and diagnostic checking reveal that the appropriate model for precipitation in this basin utilizes a nonseasonal moving average operator of order 1 at lag 3. The final model is

$$y_t = 48.46 + 0.3453 a_{t-3} + a_t,$$

with an overall standard error of 6.107. Invertibility conditions are satisfied since the absolute value of the parameter Θ_1 (-0.3453) is less than 1. Further, the Ljung-Box statistic indicates that this model accounts for nearly all of the autocorrelation between observations since the associated probabilities are quite high (0.945 at lag 6, 0.929 at lag 12, 0.986 at lag 18, and 0.822 at lag 24).

The 95% prediction interval forecast calculated for this model is 39.78 cm to 63.72 cm. Historically, the years 1934 (32.52 cm) and 1936 (34.41 cm) had annual precipitation quantities less than the lower bound of this interval. Notably, there were no years with an annual precipitation larger than the upper 95% prediction limit. This suggests that it would be quite unusual to have an annual precipitation greater than 63.72 cm.

For the surplus annual series, both the SAC and the SPAC have spikes at lag 1 and cut off. By comparison, the SPAC cuts off more abruptly, while the SAC dies down fairly quickly with damped exponential oscillation. Thus, the behavior of the SAC indicates that modelled surplus is stationary over the years examined. Interestingly, the sample autocorrelation at lag 3 is not statistically large (t_{r3} = 0.2497) despite the significant autocorrelation at this lag in the annual precipitation series. Evaluation of a nonseasonal autoregressive model of order 1 indicates that the appropriate Box-Jenkins model for this series is

$$y_t = 2.544 + 0.2687 y_{t-1} + a_t,$$

where the standard error is 2.381. The requisite stationarity conditions hold since the absolute value of ϕ_1 (0.2687) is less than 1. In addition, the probability values for the Ljung-Box statistic are high, indicating that the model adequately accounts for most of the autocorrelation between time series observations (0.722 at lag 6, 0.847 at lag 12, 0.944 at lag 18, and 0.821 at lag 24).

The 95% prediction interval of -1.64 cm to 7.70 cm forecast for the 58th observation actually exceeds possibility since it includes negative values. As mentioned earlier, the water balance surplus cannot be less than zero, so the lower limit of this interval cannot be exceeded. Historically, annual surplus quantities in 1947 (8.28 cm), 1951 (9.21 cm), 1973 (9.11 cm), 1983 (9.15 cm), and 1984 (8.58 cm) exceeded the 95% prediction interval upper limit. All of these years are noted for major floods within the Missouri basin.

The behavior of the SAC function for modelled runoff reveals that the time series is stationary. The SAC and the SPAC for modelled runoff have spikes at lag 1, and the SPAC appears to cut off more abruptly. Parameter estimation and diagnostic checking indicate that the water balance runoff is represented by a Box-Jenkins model quite similar to the model for surplus, specifically,

$$y_t = 2.436 + 0.2990 y_{t-1} + a_t,$$

with a standard error of 2.317. Stationarity conditions are satisfied since the absolute value of ϕ_1 (0.2990) is less than 1. The probability values for the Ljung-Box statistic indicate that this is an adequate final model (0.687 at lag 6, 0.811 at lag 12, 0.956 at lag 18, and 0.872 at lag 24).

A 95% prediction interval of -1.55 cm to 7.53 cm is forecast for the 58th year. Again, the limits for the 95% prediction interval exceed possibility since modelled runoff cannot be less than zero. Historically, the same years with surplus quantities greater than the upper prediction limit also had greater modelled runoff quantities: 1947 (8.05 cm); 1951 (8.53 cm); 1973 (8.79 cm); 1983 (9.16 cm); and 1984 (8.56 cm).

The SAC for annual discharge has spikes at lags 1 and 2 and cuts off after lag 2. This behavior indicates that the time series is stationary over the years studied. The SPAC has a spike at lag 1 and cuts off after lag 1. By comparison, it appears that the SPAC cuts off more abruptly, so a tentative autoregressive model of order 1 is selected. Diagnostic checking reveals that annual discharge within this basin is best described by the specific Box-Jenkins time series model

$$y_t = 2.574 + 0.4966 y_{t-1} + a_t,$$

where the overall standard error is 1.600. Stationarity conditions are met since ϕ_1 is less than 1 (0.4966). The Ljung-Box statistic probability values indicate that the model is satisfactory (0.796 at lag 6, 0.915 at lag 12, 0.938 at lag 18, and 0.857 at lag 24).

A 95% prediction interval of 1.60 cm to 7.87 cm is forecast for discharge in the 58th year. Historically, the years which exceeded these limits were 1951 (with an annual discharge of 9.16 cm), 1973 (9.24 cm), 1983 (7.92 cm), 1984 (8.37 cm), and 1987 (8.38 cm). It is interesting to note that there were no years from 1932 to 1988 with an annual discharge less than the lower prediction limit. Again, this is likely the result of a consistent base supply of streamflow from reservoir storage, groundwater sources, and snowmelt.

The behavior of the SAC calculated for the Missouri discharge fraction indicates that the annual series is nonstationary over the years examined since the SAC dies down extremely slowly. Thus, in order to continue the Box-Jenkins analysis, the series must be transformed into a stationary time series. The first differences of the original values,

$$z_t = y_t - y_{t-1},$$

are plotted in Figure 4.7. The behavior of the SAC function calculated for the first differences confirms stationarity since the function has a spike at lag 1 and cuts off. The SPAC calculated for the first differences has spikes at lags 1 and 2 and cuts off. Parameter estimates are calculated for several tentative models which indicate that the constant term μ should not be included in a parsimonious model. Diagnostic checking reveals that a nonseasonal additive moving average model of order 2 is the best model for the Missouri discharge fraction, specifically,

$$z_t = (1 - 0.5096\,B - 0.3245\,B^2)\,a_t,$$

where the overall standard error is 0.0251. Invertibility conditions for this model are satisfied since the sum of the moving average operators is less than 1, and the weights placed on the moving average operators decline as lags move further into the past. Also, the Ljung-Box statistic probability values suggest that almost all of the autocorrelation between time series values is accounted for in this model (0.796 at lag 6, 0.761 at lag 12, 0.487 at lag 18, and 0.509 at lag 24).

Since this Box-Jenkins model is based on the first differences of the original series, the model can be rewritten as

$$y_t = y_{t-1} - 0.5096\,a_{t-1} - 0.3245\,a_{t-2} + a_t.$$

The basic nature of this model is stochastic because the constant term μ is not included. Therefore, the Missouri discharge fraction is determined by values of the random shocks generated in the previous two years.

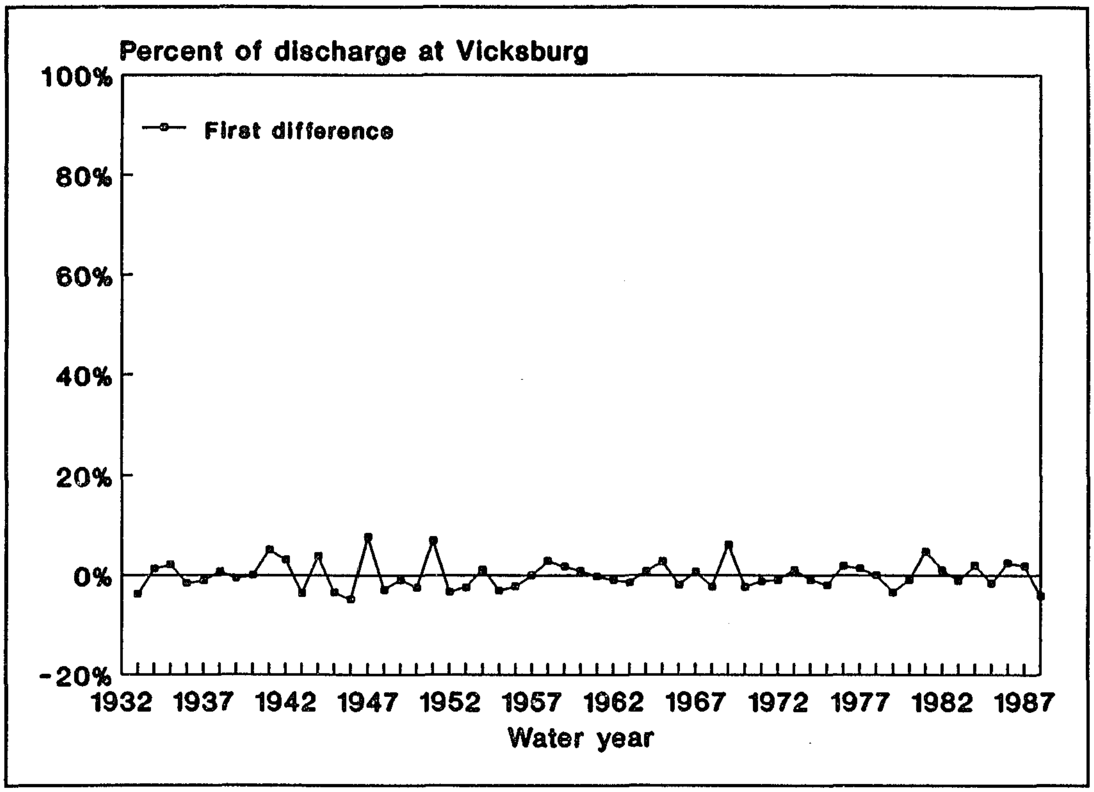

Figure 4.7. Time series graph of the first differences of the Missouri discharge fraction.

Analysis of the SAC for the water balance residuals indicates that the SAC contains no spikes. The series dies down fairly quickly, so the time series is stationary over the years examined. The SPAC contains a near spike at lag 12 ($t_{r12\,12}$ = 1.965), so a tentative autoregressive model of order 1 at lag 12 is evaluated. The parameter estimates calculated by SAS reveal a t-value for ϕ_{12} which is less than 2 (at -1.75); therefore, this operator should not be included in a parsimonious model. Next, a tentative model with no nonseasonal operator is evaluated. This model proves to be adequate, and the final model for the water balance residuals is

$$y_t = -1.647 + a_t,$$

with an overall standard error of 0.9642. There are no stationarity or invertibility conditions for this model, and the Ljung-Box probability values reveal that the model is adequate (0.921 at lag 6, 0.526 at lag 12, 0.826 at lag 18, and 0.863 at lag 24). In addition, there are no spikes in the RSPAC calculated for the Box-Jenkins model residuals.

The 95% prediction interval forecast for the water balance residual in the 58th year is -0.24 cm to 3.54 cm. In comparison, the years that produced water balance residuals outside of this interval were 1947 (-0.97 cm) and 1983 (-1.24 cm).

4.4 Summary of Findings for the Missouri Basin at Hermann

These statistical results can be translated into a summary of the annual hydroclimatology of the Missouri subbasin. Although the Box-Jenkins model for annual precipitation in the overall Mississippi basin reveals a purely random series, the model for precipitation in the Missouri subbasin indicates a three-year pattern of autocorrelation. In other words, annual precipitation values separated by a lag of three years are positively autocorrelated. It is quite likely that this three-year pattern in precipitation is the source of the three-year pattern detected in the water balance residuals for the overall Mississippi basin above Tarbert Landing. Interestingly, the Southern Oscillation is also roughly based on a three-year period (Philander 1983), and this will be discussed further in the next chapter.

The Box-Jenkins models for surplus, modelled runoff, and discharge demonstrate a positive autocorrelation between values separated by a lag of one year.

As a result, a year of high discharge is more likely to follow a year of high discharge, and a year of low discharge is more likely to follow a year of low discharge. Although the time series graphs for these variables visually appear to be increasing over the years studied, the Box-Jenkins statistical analyses reveal that the series are stationary. However, an extension of this study to include the 1993 flood in the Missouri basin could prove sufficient to produce a statistically significant long-term trend in increasing discharge.

The Missouri discharge fraction is nonstationary over the years studied which illustrates that the contribution to total discharge from the Missouri subbasin has changed significantly over time. In part, the construction of a series of dams within the subbasin appears to have reduced the discharge fraction over the years 1963 to 1967.

The Box-Jenkins model for the water balance residuals indicates that the residuals are a random, white noise series. There is a suggestion of autocorrelation at a lag of 12 in this series, yet this is not statistically significant. Except for the discharge fraction time series, the annual series examined for the Missouri subbasin are stationary over the time frame of this study. There is no statistical evidence of a significant long-term trend in precipitation, discharge, or the water budget residuals. Therefore, any determination of whether land-use changes and river management have affected discharge for the Missouri subbasin is inconclusive based on these results.

4.5 Water Balance Results for the Ohio-Tennessee Basin at Metropolis

As mentioned earlier, a significant portion of the runoff generated within the Mississippi basin originates within the Ohio-Tennessee drainage basin. The Ohio-

Tennessee subbasin comprises only 18% of the area of the Mississippi basin above Vicksburg and is less than half (39%) of the size of the Missouri subbasin; however, because of its humid climate, this relatively small drainage area produces a large quantity of streamflow.

Table 4.3 contains mean values and normality test results (W:Normal and Prob > W) calculated for the annual series which illustrate the comparatively greater quantities of precipitation and runoff generated per unit area within this drainage basin. The hypothesis that these series are samples from a normal population cannot be rejected at an α-level of 0.05.

Table 4.3. Normality Test Results for the Ohio-Tennessee Basin at Metropolis Annual Series

Annual Series	Mean cm	Std. Dev. cm	W:Normal	Prob > W
Precipitation	114.13	12.79	0.9695	0.3092
Surplus	45.52	11.17	0.9909	0.9842
Modelled runoff	45.52	11.01	0.9848	0.8643
Discharge	45.66	11.38	0.9801	0.6973
Water balance residuals	0.14	2.11	0.9870	0.9251

Figure 4.8 is a graph of the annual precipitation and water balance surplus as averaged over the area of the Ohio-Tennessee drainage basin. (Note that the y-axis scale used for precipitation is twice the magnitude of the scale used in previous precipitation figures in order to accommodate the greater annual quantities.) The mean annual precipitation averaged over this watershed is 114.13 cm (with a standard deviation of 12.79 cm). The temporal range of precipitation extends from 148.27 cm

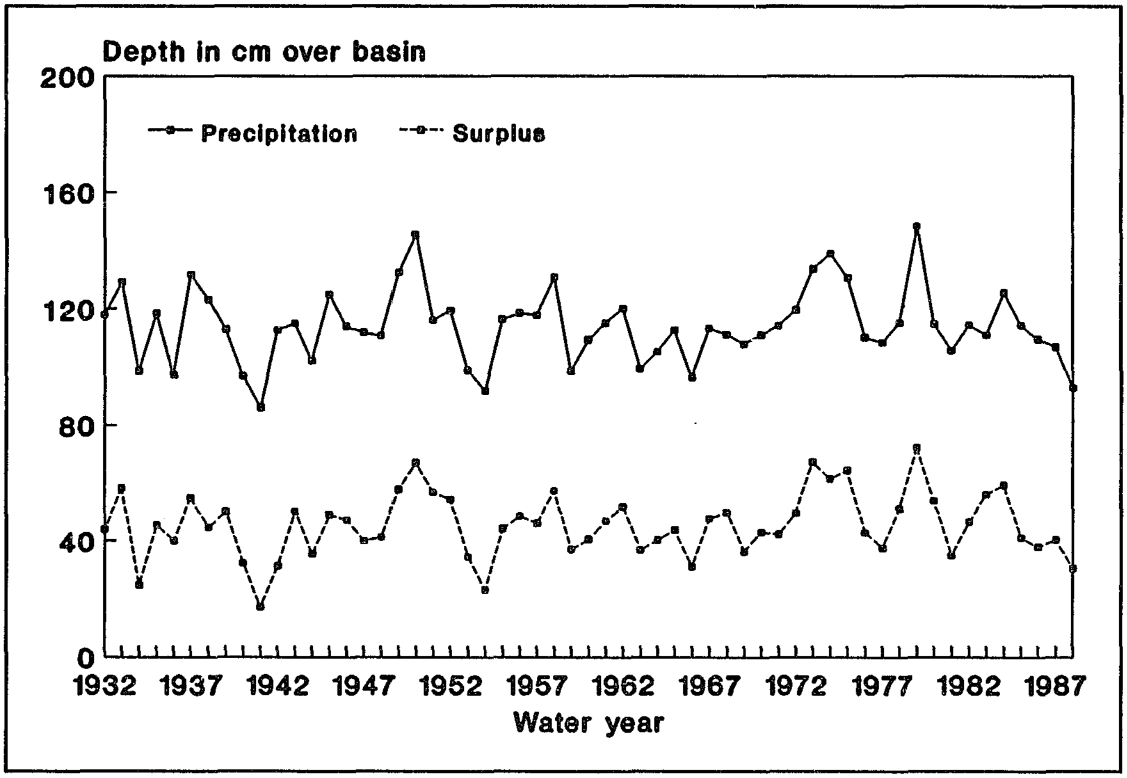

Figure 4.8. Time series graph of annual precipitation and water balance surplus for the Ohio-Tennessee basin at Metropolis.

in 1979 (130% of the mean precipitation) to 86.02 cm in 1941 (75% of the mean precipitation). It is interesting to note that even a dry year in the Ohio-Tennessee subbasin receives far more precipitation (as averaged over the drainage basin) than a flood year in the Missouri subbasin.

For the 57-year study period, annual surplus in the Ohio-Tennessee averages approximately 40% of the mean annual precipitation at 45.52 cm (with a standard deviation of 1.17 cm). Temporally, surplus ranges from 72.17 cm in 1979 (63% of the mean annual precipitation to 17.25 cm in 1941 (15% of the mean annual precipitation).

A comparison of discharge and modelled runoff is shown in Figure 4.9. (Once again, note that the y-axis scale used for modelled runoff and discharge is twice the

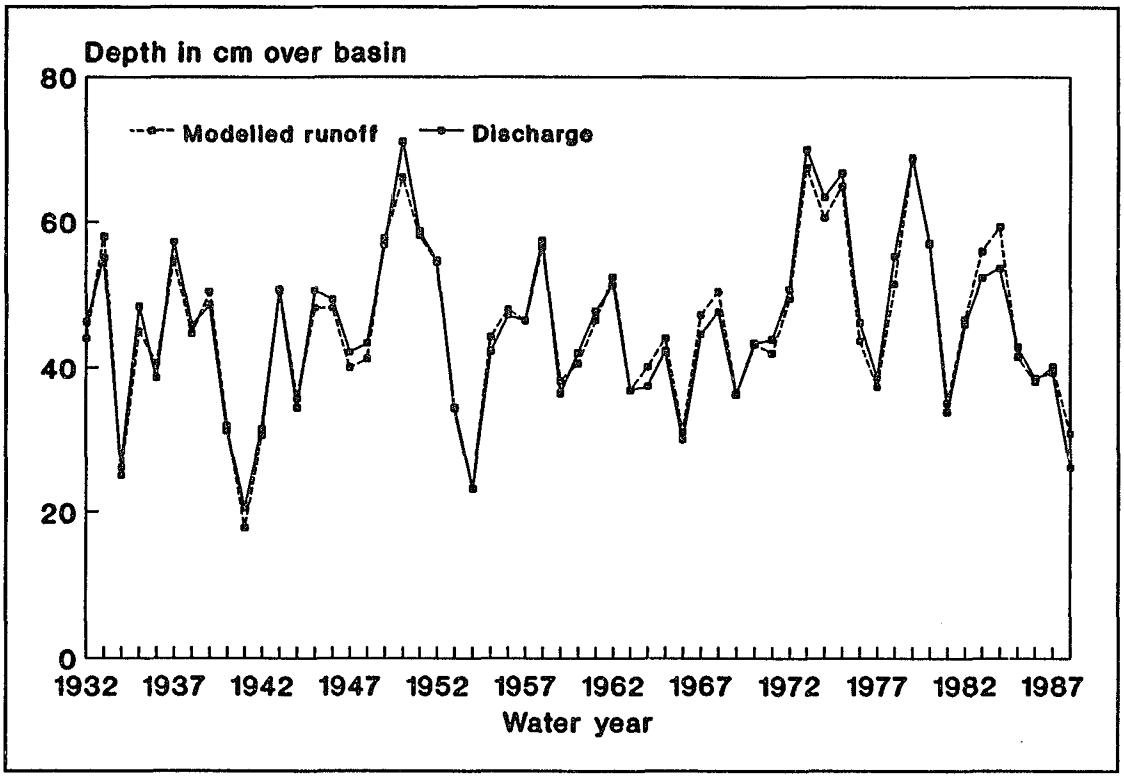

Figure 4.9. Time series graph of annual discharge and modelled runoff for the Ohio-Tennessee basin at Metropolis.

magnitude of the scale used in previous discharge graphs in order to accommodate the greater annual quantities.) On average, the water balance model for this basin only slightly underestimates discharge since the mean annual modelled runoff is 45.52 cm (0.3% underestimated). The range of modelled runoff varies from a maximum of 68.63 cm in 1979 to a minimum of 17.86 cm in 1941. Mean annual discharge for the Ohio-Tennessee basin is 45.66 cm, and the range of discharge extends from 71.08 cm in 1950 (156% of the mean discharge) to 20.54 cm in 1941 (45% of the mean discharge).

To determine whether the flow dynamics of the overall Mississippi drainage basin have changed over time, the discharge contributed by the Ohio-Tennessee basin is calculated as a fraction of the total Mississippi River discharge measured at

Vicksburg. Figure 4.10 shows the Ohio-Tennessee discharge fraction over the years 1932 to 1988. The annual volume of water generated by the Ohio-Tennessee subbasin averages nearly half (46.36%) of the annual volume of water measured at the Vicksburg gauging station and ranges from a maximum of 64.88% in 1956 to a minimum of 31.16% in 1942. It appears that the discharge fraction is not stationary over the years 1932 to 1988. The drought which occurred in the eastern U.S. from 1985 to 1988 is reflected in this graph as four years of consistently small (less than 37%) contributions to the total discharge of the Mississippi basin.

Figure 4.11 contains a time series plot of the water balance residuals for the Ohio-Tennessee basin. (Here, note that the y-axis scale for the water balance residuals is equivalent to the magnitude of the scale used in previous graphs of water balance residuals.) The average magnitude (based on absolute values of the residuals) of the water balance residuals is 1.66 cm (less than 4% of the mean annual discharge), and the residuals fluctuate around a mean value of 0.14 cm. The overall range of residual values extends from 4.99 cm in 1950 to -5.70 cm in 1984. Accordingly, the standard deviation is extremely large at 2.11 cm. It is visually apparent in this graph that the mean and variability of the water balance residuals are not constant over the 57-year period. For example, the series has a period of high variability from 1932 to 1951. Thereafter, variability is somewhat damped until 1981. Then, over a period of five years, the residual varies from 3.77 cm in 1978 to -5.70 cm in 1984, which covers nearly the full range of the water balance residuals.

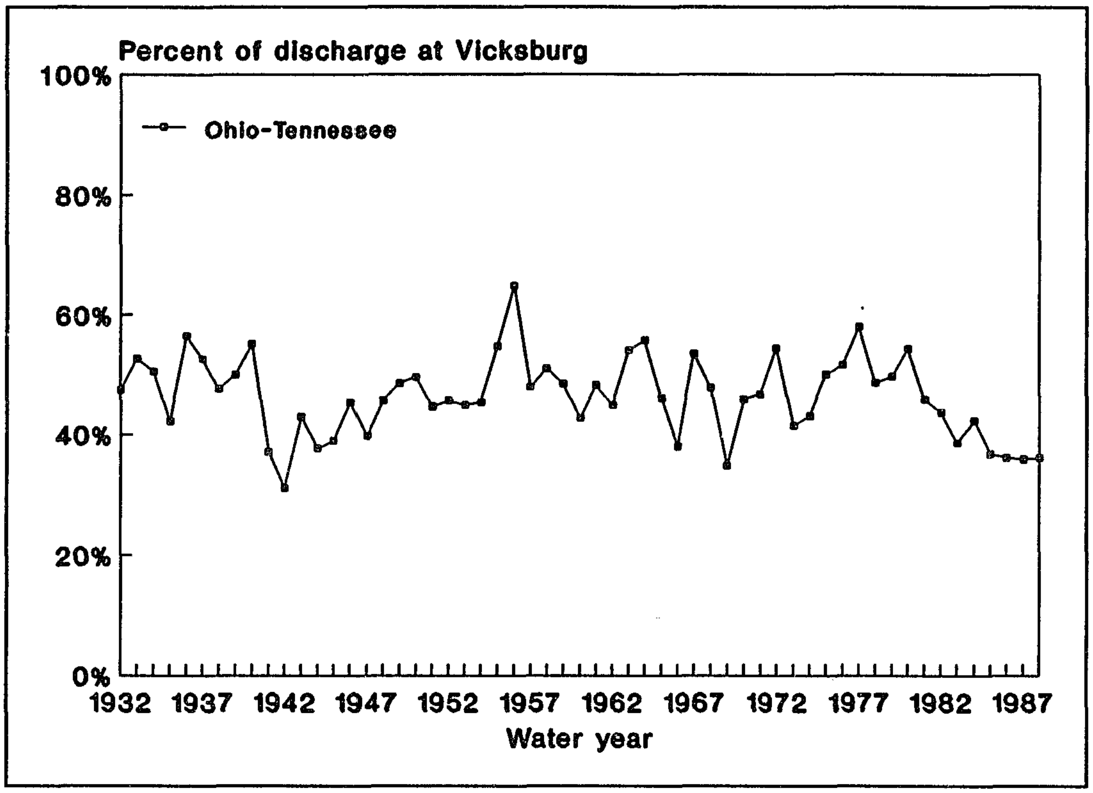

Figure 4.10. Time series graph of the discharge fraction for the Ohio-Tennessee basin at Metropolis.

Generally, the water balance model tends to overestimate measured discharge in this subbasin during the extremely dry years of the 1980's. Large negative residuals occur in 1983 (-3.61 cm), 1984 (-5.70 cm), and again in 1988 (-4.71 cm). It is possible that the Corps of Engineers released smaller quantities of discharge from the extensive reservoir system in this subbasin in order to maintain reservoir levels during these dry years. This type of response could explain these residuals. In contrast, the water balance model underestimates actual discharge to produce large positive residuals for the years 1935 (3.41 cm), 1941 (2.68 cm), 1950 (4.99 cm), 1974 (2.82 cm), and 1978 (3.77 cm). In other words, discharge was greater than the runoff climatically available in these years.

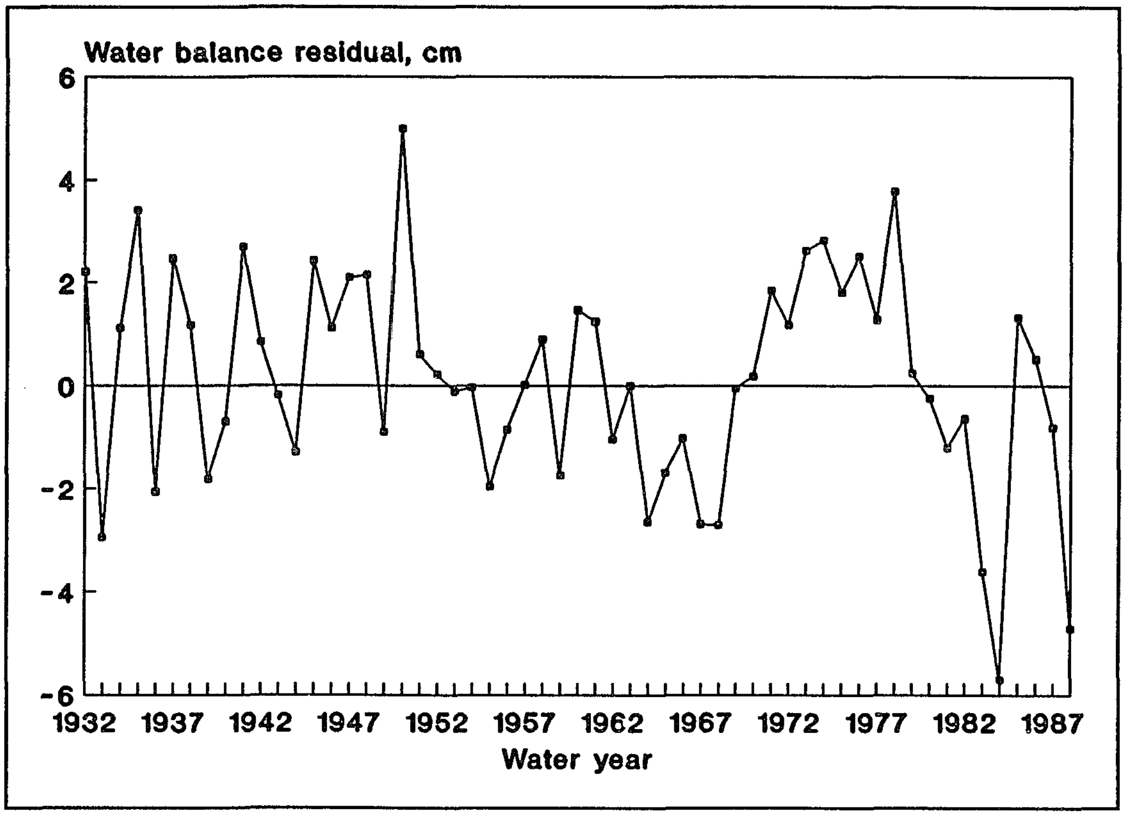

Figure 4.11. Time series graph of water balance residuals for the Ohio-Tennessee basin at Metropolis.

A plot of the cumulative water balance residuals is shown in Figure 4.12. This graph can be divided into four distinct time periods: 1932 to the early 1950's; the early 1950's to the late 1960's; the late 1960's to the late 1970's; and the late 1970's to 1988. During the first 20 years or so, the cumulative residuals are increasingly positive, indicating that measured discharge out of the basin is on average much greater than that predicted by the water balance model. Then in the 1950's, modelled runoff shifts to become greater than actual discharge for a period of years. In the late 1960's, this trend reverses such that discharge is again consistently greater than modelled runoff. In 1980, the trend reverses yet again. It is possible that revisions in the stage-

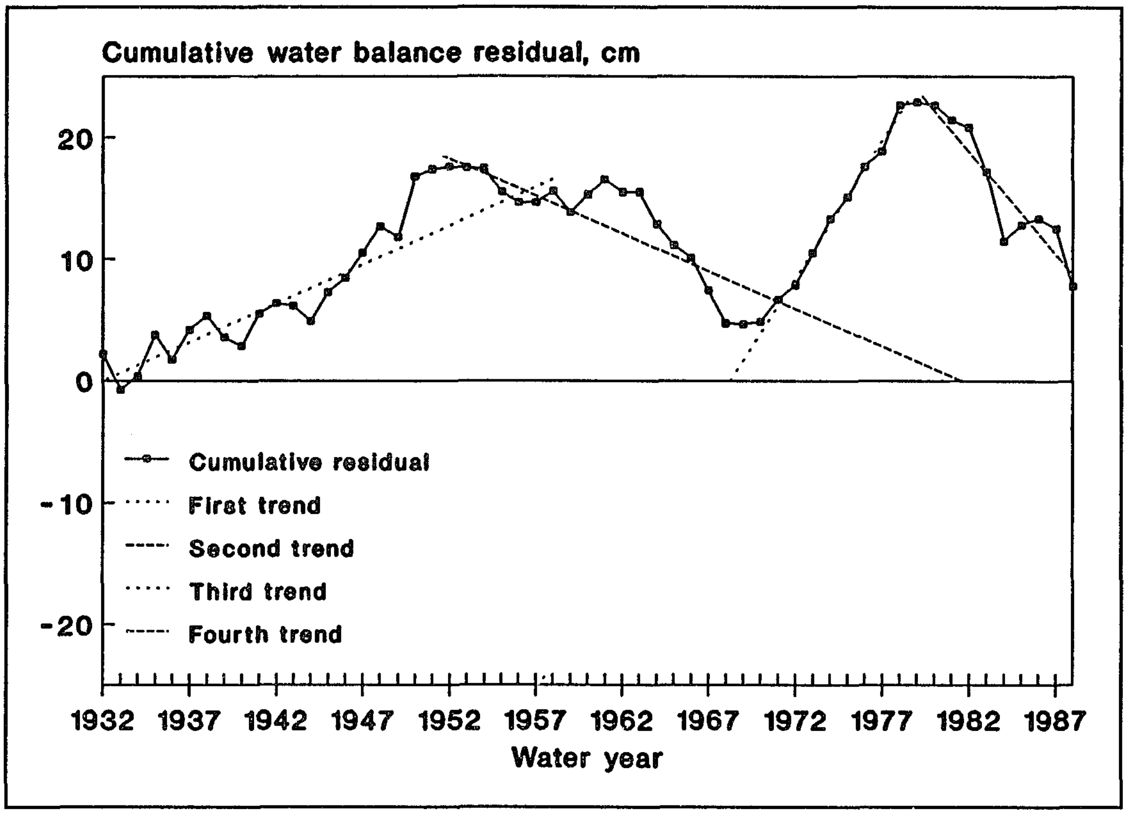

Figure 4.12. Time series graph of cumulative water balance residuals for the Ohio-Tennessee basin at Metropolis.

discharge curve have created these distinct shifts (as postulated earlier for the Tarbert Landing data), or the trends could be a result of long-term TVA management plans. Since collective impacts are involved, it is difficult to determine an exact cause. The one constant factor in this analysis is the water balance model since it accounts for climatic variability and contains none of the variability caused from human influences.

4.6 Linear Regression Results for the Ohio-Tennessee Basin at Metropolis

Of the basins selected for this study, the water balance calculated for the Ohio-Tennessee basin yields the best overall fit and the highest correlation to measured discharge. Table 4.4 contains the results of least-squares linear regression analyses for surplus versus discharge and modelled runoff versus discharge. The adjusted r^2 values are quite large for both surplus and modelled runoff indicating that the water balance model explains greater than 96% of the variability in annual discharge. Figure 4.13 is a scatter plot of modelled runoff versus discharge. In the other watersheds evaluated, the water balance is most effective at estimating discharge during average flow conditions. Notably for this basin, the water balance is just as effective at estimating extreme conditions of discharge.

The error terms are normal for these regressions. Plots of the regression errors versus discharge, predicted values, and water year reveal that the assumption of constant variance is reasonable. Also, the Durbin-Watson test statistics for surplus versus discharge (1.785) and runoff versus discharge (1.639) fall within the area for accepting the null hypothesis of zero autocorrelation of errors. Therefore, the assumptions for regression analysis are satisfied, and the regression results are assumed to be valid.

Table 4.4. Linear Regression Results of Water Balance Indices versus Discharge for the Ohio-Tennessee Basin at Metropolis

Index	F Value	Prob>F	Root MSE	Adj. r^2
Surplus*	1475.638	0.0001	2.1359	0.9634
Modelled runoff**	1562.323	0.0001	2.0487	0.9654

* $\text{Surplus}_t = 1.547 + 0.9631 \text{ discharge}_t + \epsilon_t$
** $\text{Modelled runoff}_t = 2.124 + 0.9505 \text{ discharge}_t + \epsilon_t$

4.7 Box-Jenkins Time Series Models for the Ohio-Tennessee Basin at Metropolis

The SAC calculated for annual precipitation has spikes at lags 1 and 9 and cuts off after lag 9, which indicates that annual precipitation for the Ohio-Tennessee

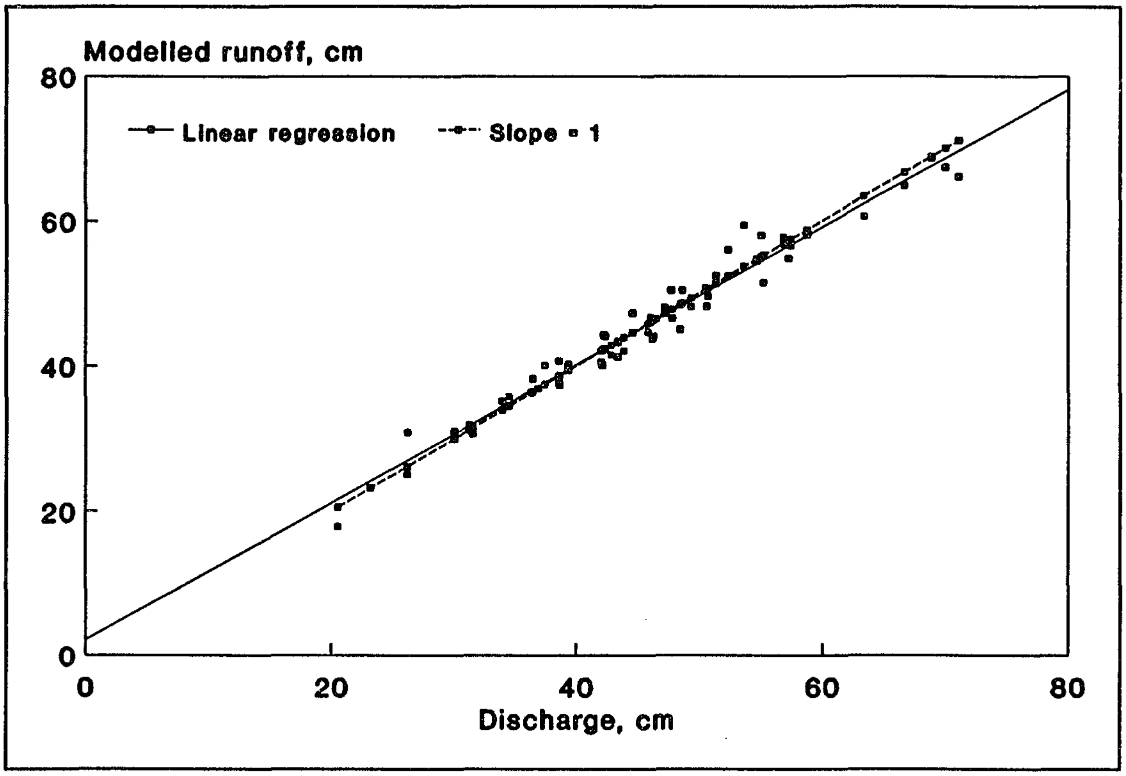

Figure 4.13. Modelled runoff versus annual discharge for the Ohio-Tennessee basin at Metropolis.

subbasin is stationary over the years studied. The SPAC has no spikes and dies down with damped sine wave oscillation. Initially, model parameters are calculated for an additive moving average model based on lags 1 and 9 as follows,

$$y_t = \mu + \left(1 - \theta_1 B - \theta_9 B^9\right)a_t .$$

The results indicate that Θ_1 should not be included in a parsimonious model since the t-value for this parameter is much less than 2 at -1.66. Evaluation of other tentative models and diagnostic checking reveal that a moving average model of order 1 at lag 9 is the best fit for this series, specifically,

$$y_t = 114.20 - 0.3360a_{t-9} + a_t,$$

which has an overall standard error of 12.18. Invertibility conditions are satisfied since the absolute value of Θ_1 (0.3360) is less than 1. Probability values calculated for the Ljung-Box statistic indicate that the model is adequate (0.494 at lag 6, 0.692 at lag 12, 0.743 at lag 18, and 0.591 at lag 24).

Based on 57 observations, a 95% prediction interval of 90.17 cm to 137.91 cm is forecast for precipitation in the 58th year. An evaluation of historical precipitation quantities shows that the annual precipitation for 1941 (86.02 cm) fell below the lower 95% prediction limit, while the years 1950, 1974, and 1979 had an annual precipitation greater than the upper 95% prediction limit (with values of 145.42 cm, 138.89 cm, and 148.27 cm, respectively).

The SAC calculated for the annual water balance surplus has a spike at lag 1 and cuts off abruptly after lag 1; therefore, the series is stationary. The sample autocorrelation at lag 9 is fairly large, yet not large enough to be statistically significant, despite the significant autocorrelation at this lag in the precipitation annual series. The SPAC has a spike at lag 1 and dies down with damped sine wave oscillation. Diagnostic checking reveals that a moving average model of order 1, specifically,

$$y_t = 45.36 + 0.3373a_{t-1} + a_t,$$

has the smallest standard error (10.66) of the tentative models examined. The necessary invertibility conditions are satisfied since the absolute value of Θ_1 (-0.3373) is less than 1. Further, probability values for the Ljung-Box statistic indicate that this is a satisfactory final model (0.529 at lag 6, 0.420 at lag 12, 0.550 at lag 18, and 0.530 at lag 24).

A 95% prediction interval of 19.91 cm to 61.68 cm is forecast for surplus in the 58th year. In comparison, the surplus calculated for the years 1941 (17.25 cm), 1950 (67.03 cm), 1973 (67.34 cm), 1975 (64.47 cm), and 1979 (72.17 cm) exceeded these limits.

The SAC computed for the water balance runoff indicates that the annual series is stationary since the function has a spike at lag 1 and cuts off. The SPAC also has a spike at lag 1. The SAC appears to cut off more abruptly, while the SPAC dies down with sine wave oscillation. Calculation of parameters and diagnostic checking for a nonseasonal moving average model of order 1 produces the final model

$$y_t = 45.36 + 0.3477 a_{t-1} + a_t,$$

which has an overall standard error of 10.46. This model satisfies the required invertibility conditions since the absolute value of Θ_1 (-0.3477) is less than 1. In addition, the probability values calculated for the Ljung-Box statistic indicate that the model is satisfactory (0.536 at lag 6, 0.416 at lag 12, 0.568 at lag 18, and 0.527 at lag 24).

This Box-Jenkins model forecasts a 95% prediction interval of 20.26 cm to 61.27 cm for runoff in the 58th year. Historically, this interval was exceeded by modelled runoff calculated for the years 1941 (17.86 cm), 1950 (66.09 cm), 1973 (67.39 cm), 1975 (64.89 cm), and 1979 (68.63 cm).

The SAC calculated for the annual discharge series is similar to the SAC for annual precipitation since the SAC has a spike at lag 1, and the sample autocorrelation at lag 9 is large (t_{r9} = 1.94). Thus, the behavior of the SAC indicates that the series is stationary over the years studied. The SPAC has a spike at lag 1 and dies down with damped sine wave oscillation. The calculation of parameter estimates for several tentative models indicates that a multiplicative moving average model of the form

$$y_t = \mu + (1 - \theta_1 B)(1 - \theta_9 B^9)a_t,$$

results in the best fit for the series. Diagnostic checking reveals that the final Box-Jenkins model for annual discharge in the Ohio-Tennessee basin is

$$y_t = 45.66 + (1 + 0.3287B)(1 - 0.3486B^9)a_t,$$

which has an overall standard error of 10.28. This model is equivalent to

$$y_t = 45.66 + 0.3287a_{t-1} - 0.3486a_{t-9} - 0.1146a_{t-10} + a_t.$$

which is based on previous values of the random shocks at lags 1, 9, and 10. Invertibility conditions are satisfied since the sum of the moving average operators is less than 1, and the weights placed on the moving average operators decline as lags move further into the past. The Ljung-Box statistic probability values indicate that nearly all of the autocorrelation between observations is accounted for by this model (0.803 at lag 6, 0.918 at lag 12, 0.875 at lag 18, and 0.812 at lag 24).

The 95% prediction interval forecast for discharge in the 58th year is 18.19 cm to 58.50 cm. Historically, annual discharge in the years 1950 (71.08 cm), 1973 (70.00 cm), 1974 (63.40 cm), 1975 (66.69 cm), and 1979 (68.88 cm) was greater than the upper prediction limit. There were no years with an annual discharge less than the lower prediction limit. Again, this is likely the result of a consistent base flow due to groundwater supplies and reservoir storage within the drainage basin.

The SAC calculated for the Ohio-Tennessee discharge fraction dies down extremely slowly, which indicates that this time series is nonstationary over the years examined. It appears that a first differencing transformation results in a stationary series as shown in Figure 4.14. The SAC for the first differences has spikes at lags 1 and 2 and cuts off after lag 2. An examination of the SPAC indicates that the SPAC has spikes at lags 1 and 2 and cuts off after lag 2. Since it is not immediately obvious which function cuts off more abruptly, several tentative models utilizing moving average operators and autoregressive operators are examined. From these estimates, diagnostic checking reveals that a multiplicative moving average model of order 2 has the lowest standard error. Therefore, the final Box-Jenkins model for the Ohio-Tennessee discharge fraction is

$$z_t = (1 - 0.4392\,B)(1 - 0.3627\,B^2)\,a_t,$$

or equivalently,

$$y_t = y_{t-1} - 0.4392\,a_{t-1} - 0.3627\,a_{t-2} + 0.1593\,a_{t-3} + a_t.$$

The overall standard error for this model is 0.0637, and the probabilities calculated for the Ljung-Box statistic indicate that the model is adequate (0.657 at lag 6, 0.596 at lag 12, 0.674 at lag 18, and 0.801 at lag 24). The invertibility conditions required for a second-order moving average model are satisfied. Therefore, the Ohio-Tennessee discharge fraction is determined by values of the random shocks generated in the

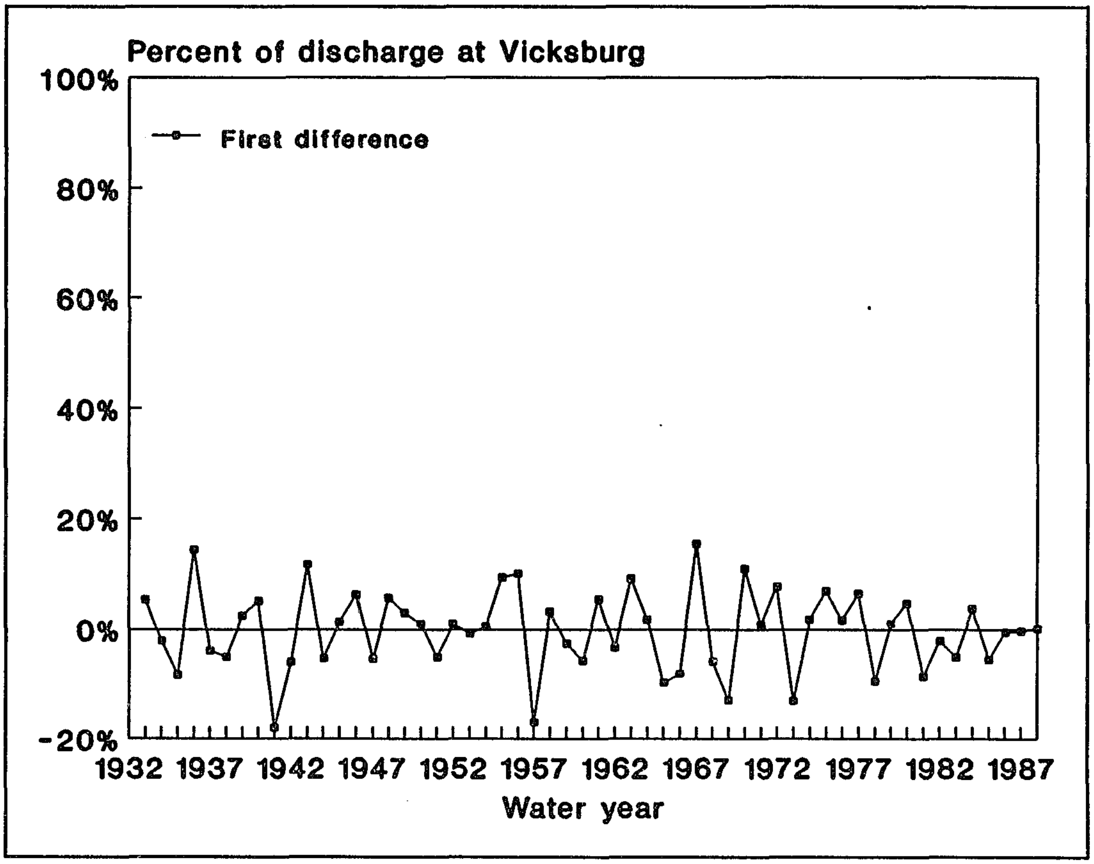

Figure 4.14. Time series graph of the first differences of the Ohio-Tennessee discharge fraction.

previous three years. The basic nature of this model is stochastic because the constant term μ is not included in the model.

An analysis of the SAC calculated for the water balance residuals reveals that the SAC dies down extremely slowly rather than cutting off or dying down fairly quickly. This type of behavior indicates that the water balance residuals are not stationary over the years examined. The first differences of the original values are plotted in Figure 4.15. The SAC function calculated for the first differences has a spike at lag 1 and at lag 14. Since the SAC cuts off quickly, it appears that first

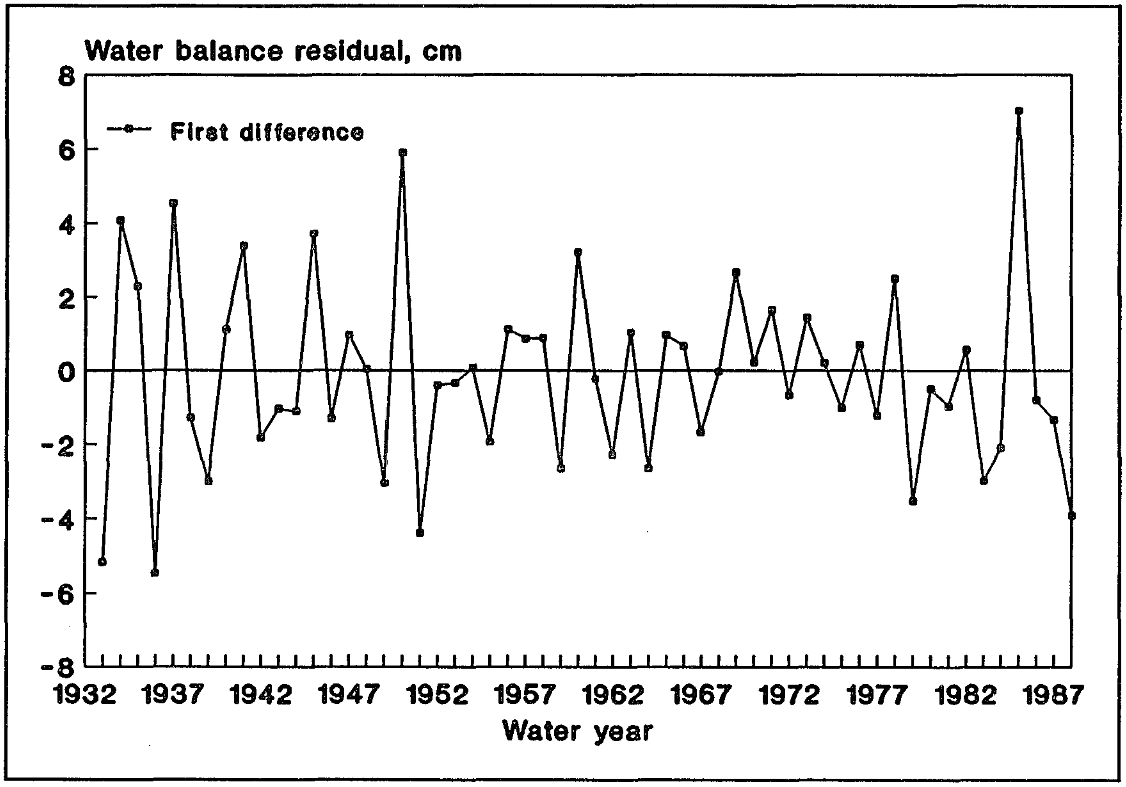

Figure 4.15. Time series graph of the first differences of the water balance residuals for the Ohio-Tennessee basin at Metropolis.

differencing has produced a stationary series. The SPAC for the transformed data has spikes at lags 1 and 2 and dies down. Calculation of model parameters and diagnostic checking for several tentative models indicates that the t-value for the constant term μ is much less than 2, which indicates that μ is not significant and should not be included in a parsimonious model. Further, the t-value for a moving average parameter at lag 14 is much less than 2 (1.68) in a multiplicative model of the form

$$z_t = \left(1 - \theta_1 B\right)\left(1 - \theta_{14} B^{14}\right) a_t .$$

However, analysis of a tentative moving average model of order 1 has a spike at lag 14 (2.172) in the RSPAC function. Thus, in order to account for the autocorrelation at lag 14, the best model for this series is a multiplicative moving average model with operators at lags 1 and 14,

$$z_t = (1 - 0.6608B)(1 - 0.2884B^{14})a_t,$$

where

$$z_t = y_t - y_{t-1}$$

for t = 2, 3, 4, ..., n. This model is equivalent to

$$y_t = y_{t-1} - 0.6608a_{t-1} - 0.2884a_{t-14} + 0.1906a_{t-15} + a_t.$$

which is based on previous values of the random shocks at lags 1, 14, and 15. The standard error (2.097) calculated for this model is the lowest of the tentative models evaluated. Invertibility conditions hold since the sum of the values of the parameters is less than 1, and the weight of the parameters declines as lags grow longer. The Ljung-Box statistic probability values indicate that the model is sufficiently adequate (0.518 at lag 6, 0.782 at lag 12, 0.654 at lag 18, and 0.899 at lag 24). As a final check, there are no spikes in the RSAC and RSPAC functions calculated for this model.

The 95% prediction interval forecast for the 58th year is -6.31 cm to 1.91 cm. A comparison of historic values to this interval reveal that a number of years had water balance residuals greater than the upper 95% prediction interval: 2.21 in 1932; 3.41 cm in 1935; 2.46 in 1937; 2.68 cm in 1941; 2.42 in 1945; 2.09 in 1947; 2.14 in 1948; 4.99 cm in 1950; 2.61 in 1973; 2.82 cm in 1974; 2.50 in 1976; and 3.77 cm in 1978. Thus, in each of these years, more discharge was measured than climatically accounted for by the water balance.

4.8 Summary of Findings for the Ohio-Tennessee Basin at Metropolis

The Box-Jenkins statistical results for this drainage basin can be interpreted within a physical context as follows. The time series model for precipitation indicates that a nine-year pattern exists in annual precipitation averaged over the Ohio-Tennessee basin. Comparatively, this pattern occurs on a longer time frame than the three-year pattern seen in the Missouri basin.

A nine-year autocorrelation is not significant in the Box-Jenkins model for the water balance surplus. Rather, the surplus series displays hydrologic persistence and is dependent on the random shock which occurs in the previous year. The Box-Jenkins model derived for runoff is quite similar to the surplus model and provides a better estimate of annual discharge in the Ohio-Tennessee subbasin.

A nine-year autocorrelation is significant in the Box-Jenkins model for annual discharge which presumably originates from the nine-year pattern detected in the precipitation annual series. In addition, hydrologic persistence results in a dependence on previous year values of the random shock term. The final model depends on moving averages of the random shock at one-, nine-, and ten-year lags.

The Ohio-Tennessee discharge fraction is nonstationary over the years studied which illustrates that the contribution to total discharge from this subbasin has changed significantly over time. The quantity of discharge produced by this subbasin is dependent on large-scale climate patterns, yet this quantity could also be influenced by reservoir levels within the subbasin and TVA management policies.

The Ohio-Tennessee subbasin provides the most interesting case in regard to nonseasonal time series analysis of the water balance residuals since the original time series values are nonstationary. It is quite obvious that the actual runoff from this basin is considerably different from the climatic runoff estimated by the water balance model. Thus, the quantity of runoff which can be attributed to factors other than climate (land-use changes, river management, groundwater supplies, reservoir storage, etc.) appears to have changed significantly over the years studied. A first difference transformation effectively models the trend in the original observations since it produces a stationary series. The trend in the original time series values is stochastic. The fact that a deterministic trend is not apparent in the water balance residuals reveals that there is no long-term increase or decrease in the time series over the years examined.

4.10 Chapter Summary

This chapter has presented statistical data, time series graphs, linear regression analyses, and Box-Jenkins models to analyze the contribution of two major subbasins to the overall Mississippi drainage basin. For the Ohio-Tennessee basin, the water balance indices of surplus and modelled runoff estimate annual discharge with extremely successful results since the model is able to account for greater than 96% of the annual variance in streamflow. For the Missouri basin, the water balance model accounts for approximately 87% of the annual variance in streamflow.

In regard to the Box-Jenkins time series analyses, precipitation is estimated by a moving average time series in both basins. In the Missouri watershed, the moving average parameter is based on a lag of three years; in the Ohio-Tennessee, the moving average parameter occurs at a lag of nine years. Surplus, modelled runoff, and discharge time series models indicate autocorrelation based on a one-year moving average or autoregressive component. In addition, the discharge series for the Ohio-Tennessee reflect an autocorrelation at lags 9 and 10.

Together, the Ohio-Tennessee and the Missouri subbasins account for nearly 60% of the total streamflow generated within the overall Mississippi watershed. The size of the overall Mississippi basin is so great that even large-scale weather patterns cannot affect the entire drainage basin at the same time. For example, it is possible to have drought conditions in one region of the Mississippi basin while flood conditions exist in another portion of the watershed. As a result, the fractions of total discharge produced by the Missouri and Ohio-Tennessee subbasins are not stationary over the years 1932 to 1988.

The water balance residuals are purely random for the Missouri basin, while the residuals for the Ohio-Tennessee basin demonstrate changes in the relationship between discharge and modelled runoff over time. It is possible that these changes can be attributed to river and reservoir management by TVA and the Corps of Engineers.

CHAPTER 5

SUMMARY AND CONCLUSIONS

5.0 Chapter Objectives

The main objective of this chapter is to discuss and summarize the results from previous chapters. This study is defined by several questions concerning the accuracy of a water balance of the Mississippi River basin and the effects of human activities on streamflow. These questions are restated and discussed in light of the statistical results presented in Chapters 3 and 4. Limitations in the study design are mentioned, and policy considerations are briefly reviewed. Conclusions are drawn from the statistical results, and suggestions are included for future research.

5.1 Discussion of Results and Summary

How well does the water balance model predict annual streamflow for the Mississippi basin and its subbasins? The Thornthwaite water balance model is surprisingly accurate in estimating annual discharge from the Mississippi basin and its subbasins. Despite the large spatial area of the overall drainage basin, the calculation of water balance indices and the aggregation of data for areas as small as climate divisions proves to be quite effective in modelling annual discharge. However, it is important to note that the factors which affect the accuracy of the water balance (e.g., data errors, model errors, groundwater recharge, and land-use changes and river regulation) could offset each other and still produce a good fit for the overall basin. As a result, the information contained in the water balance residuals regarding the effects of human actions may be masked by competing effects across the large geographic area of the overall basin. Therefore, the results obtained in this study are limited by the spatial scale of large drainage basins.

Further, the results of this study are limited by the temporal scale of a water year. Although the water balance model is extremely accurate on an annual basis, the model breaks down on a seasonal basis due to problems in estimating potential evapotranspiration. The Thornthwaite equation tends to underestimate potential evapotranspiration in winter and overestimate potential evapotranspiration in summer. Consequently, modelled runoff is overestimated in winter and underestimated in summer. Therefore, a seasonal analysis using the Thornthwaite potential evapotranspiration would likely produce larger water balance residuals and compound the difficulties of residual analysis.

Table 5.1 contains a summary of the mean annual quantities of discharge, modelled runoff, and the water balance residuals calculated for each watershed. Notably, the mean annual water balance residuals (based on the absolute values of the residuals) calculated for the two subbasins are larger than the mean residuals calculated for the overall basin. This suggests that contemporaneous effects within the Mississippi drainage basin average out to yield smaller residuals for the overall basin. The net result is that the water balance explains approximately 95% of the annual discharge for the overall Mississippi basin.

The magnitudes of the mean annual water balance residuals calculated for the two subbasins are similar; however, the mean annual discharge for the Ohio-Tennessee subbasin is an order of magnitude higher than the discharge for the Missouri subbasin.

Thus, the water balance model is considerably more accurate for the Ohio-Tennessee subbasin. Least-squares linear regression results show that the modelled runoff explains over 96% of the variability in annual discharge for the Ohio-Tennessee subbasin as compared to 87% for the Missouri subbasin. Presumably, the water balance model is more accurate in estimating discharge for the Ohio-Tennessee drainage basin due to a humid climate and a smaller spatial area. The water balance model is least successful in the Missouri subbasin because of orographic precipitation that falls mostly as snow in the Rocky Mountains.

Table 5.1. Effectiveness of the Water Balance Model for the Mississippi Basin and Subbasins

Drainage Basin	Mean Annual Discharge cm	Mean Annual Modelled Runoff cm	Mean Annual Residual cm	Variability Explained by Modelled Runoff
Mississippi (Tarbert Landing)	18.82	19.12	0.91	95.22%
Mississippi (Vicksburg)	18.08	18.23	0.95	94.97%
Missouri	5.14	3.50	1.74	86.70%
Ohio-Tennessee	45.66	45.52	1.66	96.54%

Has annual discharge from the overall basin and its subbasins changed over time? The Box-Jenkins time series analysis indicates that annual discharge within the Mississippi, the Missouri, and the Ohio-Tennessee drainage basins has not changed significantly over the years 1932 to 1988. Table 5.2 contains a summary of the Box-Jenkins time series models developed for each drainage basin. The Box-Jenkins models for discharge in the Mississippi and Missouri basins display significant autocorrelation based on a one-year lag due to hydrologic persistence. A longer pattern of autocorrelation is apparent within the Ohio-Tennessee subbasin where a nine-year lag is statistically significant in the annual discharge time series. Most likely this nine-year pattern of autocorrelation results from the nine-year pattern detected in Ohio-Tennessee subbasin annual precipitation.

The Missouri subbasin discharge time series visually appears to have a long-term trend toward increasing discharge; however, a long-term trend is not statistically significant. A least-squares regression of annual discharge versus water year for the subbasin shows an increasing slope that is statistically significant; however, this result is tempered by the fact that the time series observations for annual Missouri discharge are strongly autocorrelated at lag 1, and a linear regression model does not account for serial autocorrelation. Incorporating data for the 1993 flood into this study could prove sufficient to create a statistically significant trend toward increasing discharge.

The annual discharge fractions contributed by the Missouri and the Ohio-Tennessee subbasins are nonstationary over the years studied. Some of these changes may be attributable to human actions as illustrated by the Missouri subbasin when reservoir filling seems to have reduced the discharge fraction over a period of years. Most likely the main reason for changes in the portion of total flow generated by each subbasin is that climate patterns have not been consistent from region to region and from year to year within the overall Mississippi basin. Because of the large spatial area of the basin, even large-scale weather patterns cannot extend across the entire drainage basin. For example, it is extremely rare for drought to affect the entire Mississippi watershed at the same time. Therefore, it is fairly common for one region to experience drought conditions while another region experiences floods or wetter conditions.

Table 5.2. Summary of Nonseasonal Box-Jenkins Models for Annual Series for the Mississippi Basin and Subbasins

Annual Series and Drainage Basin	Stationary?	Box-Jenkins Time Series Model
Precipitation		
Mississippi (Tarbert Landing)	Yes	White noise
Mississippi (Vicksburg)	Yes	White noise
Missouri	Yes	MA (3)
Ohio-Tennessee	Yes	MA (9)
Surplus		
Mississippi (Tarbert Landing)	Yes	MA (1)
Mississippi (Vicksburg)	Yes	AR (1)
Missouri	Yes	AR (1)
Ohio-Tennessee	Yes	MA (1)
Modelled runoff		
Mississippi (Tarbert Landing)	Yes	AR (1)
Mississippi (Vicksburg)	Yes	AR (1)
Missouri	Yes	AR (1)
Ohio-Tennessee	Yes	MA (1)
Discharge		
Mississippi (Tarbert Landing)	Yes	MA (1)
Mississippi (Vicksburg)	Yes	AR (1)
Missouri	Yes	AR (1)
Ohio-Tennessee	Yes	MA (9)
Water balance residuals		
Mississippi (Tarbert Landing)	Yes	MA (3)
Mississippi (Vicksburg)	Yes	White noise
Missouri	Yes	White noise
Ohio-Tennessee	No	ARIMA (0,1,14)
Discharge fraction		
Missouri	No	ARIMA (0,1,2)
Ohio-Tennessee	No	ARIMA (0,1,3)

This study has not considered extreme discharge events (i.e., floods or low flow events), so these results are limited to an annual basis. Further, this study considers discharge generated from extremely large watersheds and yields no information about discharge generated on a local area scale.

Has annual precipitation over the basin and its subbasins changed over time? The Box-Jenkins models developed for annual precipitation within the Mississippi basin suggest that precipitation has not changed significantly over the years 1932 to 1988. A three-year pattern of autocorrelation is detected in annual precipitation for the Missouri subbasin, and a nine-year pattern of autocorrelation is evident in annual precipitation for the Ohio-Tennessee subbasin. In general, long term cycles such as these can be attributed to large-scale aspects of the atmospheric circulation (Kahya and Dracup 1993). Interestingly, these patterns are obscured by averaging the annual precipitation for the overall area of the Mississippi basin so that the annual precipitation for the overall drainage area appears purely random. From this, it can be assumed that competing factors average out over the large geographical size of the watershed. As a result, the smaller drainage basins potentially yield more information about patterns of autocorrelation within time series observations.

Have human modifications (land-use changes and river regulation) affected annual discharge in the Mississippi basin and its subbasins over time? The water balance residuals for the Mississippi basin at Tarbert Landing display a three-year pattern of autocorrelation that seems to result from the three-year pattern of autocorrelation detected in the Missouri subbasin annual precipitation. The water balance residuals for the Mississippi basin at Vicksburg and the Missouri basin at Hermann are purely random white noise time series. Thus, it appears that any effects on annual discharge caused by human actions have been either inconsequential or masked by competing factors.

The water balance residuals for the Ohio-Tennessee basin at Metropolis are nonstationary over the years examined. The fact that the residuals have changed significantly over the years 1932 to 1988 suggests that annual discharge has been affected by human modifications within the subbasin. The Box-Jenkins model developed for this series depends on the previous year value plus a fraction of the random shocks calculated for lags of 1, 14, and 15 years. It is important to point out that the length of the time series is relatively short (57 years) for a lag of this length.

Table 5.3 shows the years with the largest water balance residuals for each drainage basin. Interestingly, some of these years are El Niño-Southern Oscillation (ENSO) years. The years 1941 and 1983 produced the major ENSO events of this century (Kahya and Dracup 1993) with heavy rains in the West, Southwest, and Colorado River basin. It is possible that ENSO effects are another factor to be considered in the analysis of the water balance residuals.

The water balance residuals represent errors in discharge measurement, errors in the water balance model, inaccuracies created by differences between drainage areas and composite climate division areas, factors in the overall water balance not accounted for by the water balance model, and changes in discharge caused by river management and land-use changes. This study was initiated with the assumption that changes in discharge caused by human actions would be considerably larger than the other sources of error. Therefore, it was assumed that it would be possible to clearly distinguish and identify changes caused by human modification. However, it seems likely that the other sources of error are generally of the same magnitude so that it is not possible to isolate only human-induced effects on discharge. Even if the water balance model could be improved considerably (by improving the estimate of potential evapotranspiration, adjusting lag time, etc.), the magnitude of other errors could remain sufficiently large to mask any effects of river regulation and land-use changes. Therefore, the study results are largely inconclusive regarding a change in discharge due to human modification for the overall Mississippi basin. However, annual discharge within the Ohio-Tennessee subbasin appears to have been influenced by human actions over the years 1932 to 1988.

5.2 Policy Considerations

Although a definitive statement about regulation and management of the Mississippi River was not the purpose of this study, policy considerations are briefly reviewed. Changnon (1987) discussed key questions to be answered in regard to the potential effects of climate change on water resources. Similar questions can be asked in regard to the effects of human impacts on the Mississippi River flow regime.

What type of effect is likely to occur? This study suggests that the effects of human impacts on the annual discharge of the Ohio River are significant. However, this study also suggests that the effects of human modification tend to cancel out when averaged over the entire Mississippi basin. It is difficult to extrapolate these results to shorter time periods and geographically smaller watersheds because of the spatial and temporal scale limitations in this study.

Table 5.3. Water Years Producing the Largest and Smallest Water Balance Residuals in the Mississippi Basin and Subbasins

Drainage Basin	Largest Positive Water Balance Residuals cm	Largest Negative Water Balance Residuals cm
Mississippi (Tarbert Landing)	1963 1950 1934 1981 1956	1947 1973 1984 1983 1968
Mississippi (Vicksburg)	1948 1934 1963 1950 1974	1983 1984 1973 1942 1947
Missouri	1948 1950 1949 1981 1943	1983 1947 1984 1942 1962
Ohio-Tennessee	1950 1978 1935 1974 1941	1984 1988 1983 1933 1968

Will the effects extend beyond the stationarity assumptions typically used in water resources design? The results from this study suggest that the effects of human modification may be sufficient to create conditions of nonstationarity over time.

Can the effects be predicted? The Box-Jenkins models developed here can be used for forecasting future annual quantities, although the predictions are limited by the length of the time series and the problems in isolating the effects of human actions.

What are the probabilities and confidence levels that the effects will occur? Because of the noise inherent in the model and the input data, it will be difficult to determine probabilities and confidence levels.

How serious are the effects? The results of this study suggest that the effects are not yet serious on an annual basis. However, if human impacts played a role in exacerbating the 1993 flood, the potential effects are quite serious.

What are the potential solutions? Further research is required to better define the problem and answer the preceding questions before specific solutions can be offered.

It would be unwise to base policy decisions strictly upon this study due to the limitations inherent in the study design. On an annual basis, the Corps of Engineers may be justified in stating that they have had no effect on the Mississippi River's discharge; however, annual discharge in the Ohio-Tennessee basin appears to have been significantly affected by human-induced changes over the years 1932 to 1988.

5.3 Conclusions

The conclusions determined from this study are summarized as follows:

1. The water balance is effective in estimating annual discharge within the Mississippi basin and the Missouri and Ohio-Tennessee subbasins. The results

of this study reveal that the water balance model explains as much as 96% of the variability in annual discharge.

2. There is no evidence of statistically significant changes or long-term trends in annual precipitation, water balance surplus, modelled runoff, and discharge in the Mississippi basin and the Missouri and Ohio-Tennessee subbasins. This study suggests no evidence of climate change on an annual basis for the years 1932 to 1988.

3. The annual discharge fractions for the Missouri and Ohio-Tennessee subbasins have changed significantly over the years 1932 to 1988. This can be attributed to the large geographic scale of the overall drainage basin with respect to large-scale atmospheric circulation patterns.

4. The relationship between annual discharge and modelled runoff has not changed in the Mississippi and Missouri basins; however, the relationship between annual discharge and modelled runoff has changed significantly in the Ohio-Tennessee subbasin over the years 1932 to 1988. These changes can possibly be attributed to river management and land-use changes within the subbasin.

5.4 Suggestions for Future Research

Future research should attempt to explain specific water balance residuals in regard to reservoir construction, storage capacity, management, revised discharge rating curves, and groundwater withdrawal and recharge quantities. This detailed analysis would provide additional insight to the effects of human impacts on annual discharge within the Mississippi basin.

The period of analysis should be extended to include 1993 data in future work. There is a visual suggestion of increasing annual discharge, water balance surplus, and modelled runoff in the Missouri subbasin which could prove to be significant with the addition of the 1993 flood to the period of study. In addition, it would be interesting to evaluate the temperature and evapotranspiration time series developed for the Mississippi basin and its subbasins.

As discussed earlier, a seasonal evaluation using the Thornthwaite equations for potential evapotranspiration would be difficult because of inaccuracies in estimating potential evapotranspiration on a shorter time scale. Future research could utilize a different method to estimate potential evapotranspiration. This would allow floods within the Mississippi basin and subbasins to be evaluated since they typically occur on shorter time scales.

Further, time series data for smaller watersheds could be evaluated to determine whether significant changes are easier to detect on a smaller geographic scale. This would provide information in regard to the local effects of human activities on discharge.

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DOCTORAL EXAMINATION AND DISSERTATION REPORT

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Major Field: Geography

Title of Dissertation: A Water Balance Evaluation of the Effects of Climate Variability and Human Modification on the Flow Regime of the Mississippi River: 1932-1988

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